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Petrography and mineral compositions of eclogites from the Koidu Kimberlite Complex, Sierra Leone

Agnes T. Fung[1] and Stephen E. Haggerty

Department of Geology, University of Massachusetts-Amherst

Abstract. The origins of eclogite (clinopyroxene + garnet) and the relative proportions of eclogite in the upper mantle are issues of considerable uncertainty and debate that bear upon the chemical and physical dynamics of petrogenesis, recycling, and remote sensing interpretation. Forty-one upper mantle eclogites from the Koidu Kimberlite Complex, Sierra Leone, were selected for detailed petrographic and chemical examination to bolster an earlier database with a view to the identification of protoliths and possible source regions of eclogite origin. On the basis of MgO contents in garnets, eclogites are divided into a high-MgO suite and a low-MgO suite. High-MgO eclogites contain pyrope (16.5–20.2 wt % MgO), with an average garnet composition of $Pyr_{65}Alm_{20}Gross_{15}$. Almandine and grossular (5.3–13.2 wt % MgO) are present in low-MgO eclogites and vary from $Pyr_{20}Alm_{60}Gross_{20}$ to $Pyr_{35}Alm_{30}Gross_{35}$. Pyroxenes in high-MgO eclogites are diopsidic ($Jd_{11}Di_{89}$-$Jd_{26}Di_{74}$); those in low-MgO eclogites range from jadeitic-diopside to omphacite ($Jd_{20}Di_{80}$-$Jd_{48}Di_{52}$). Oriented apatite crystals in garnet and clinopyroxene are interpreted to be products of exsolution and, coupled with coexisting rutile, imply that garnet is a major repository for P, Cl, F, OH, and Ti in the upper mantle. Reconstructed bulk compositions of high- (15.0–18.9 wt %) and low-MgO (7.1–12.2 wt %) eclogites are distinct, and major elements in these xenoliths broadly resemble basalts, picrites, and komatiites. Most high-MgO eclogites equilibrated at 1080°C at 4.7 GPa to 1130°C at 5.2 GPa, whereas most low-MgO eclogites cluster at 880°C at 3.3 GPa to 930°C at 3.8 GPa. Estimated PT and depths of origin of the Koidu eclogites imply that high-MgO eclogites are asthenospheric, low-MgO eclogites are lithospheric, and both are likely products of plume activity. Diamondiferous eclogites, worldwide, have characteristic Na, K, Ti, and ^{IV}Al (Si) in garnet-clinopyroxene pairs that point to distinctive source regions and petrogenesis; these chemical variables may be applied to diamond exploration and evaluation.

Introduction

Earth's upper mantle is defined as the region between the Mohorovicic discontinuity and the transition zone [*Dawson*, 1984; *Anderson*, 1989]. Seismic *P* wave velocities of the upper mantle typically range from 8.0 to 8.2 km/s [*Anderson*, 1989], which limits the petrology of the mantle at high pressure and temperature to combinations of only a few minerals: olivine, clinopyroxene, orthopyroxene, spinel, and garnet. These minerals, with some modal and chemical variations, are the major constituents of eclogite (garnet + clinopyroxene), pyroxenite (clinopyroxene + orthopyroxene ± garnet), lherzolite (olivine + clinopyroxene + orthopyroxene ± garnet), and harzburgite (olivine + orthopyroxene). These rocks exist as deep-seated, ultramafic bodies in the mantle and may be fragmented and transported to Earth's surface as xenoliths by volcanic activity. Xenoliths therefore provide invaluable insights to temperature, pressure, crystallization, fractionation, exsolution, phase changes, partial melting, carbonation, hydration, and oxidation in the upper mantle [*Sobolev*, 1977; *Dawson*, 1980; *Smith*, 1988].

The majority of mantle xenoliths are entrained in kimberlites, lamproites, and alkali basalts [*Dawson*, 1980; *Nixon*, 1987; *Jaques et al.*, 1989]. All kimberlites and lamproites contain variable amounts of ultramafic xenoliths, xenocrysts, and two suites of diamonds: a peridotitic suite and an eclogitic suite. Mineral inclusions in ultradeep (>300 km) diamonds, however, are almost exclusively eclogitic [*Moore and Gurney*, 1985]. The abundance of eclogitic suite diamonds incorporated in kimberlites and lamproites implies that eclogite is possibly a significant component of the deeper parts of the upper mantle and thus may provide important clues to the composition and evolution of Earth's mantle.

Eclogites are either igneous or metamorphic, mafic to ultramafic, garnet + clinopyroxene rocks that are formed at high pressure and temperature. These rocks are generally coarse-grained, equigranular, and basaltic to komatiitic in composition. Common accessory minerals include kyanite, corundum, graphite, diamond, quartz-coesite, and sanidine. Although eclogitic suite diamonds have been widely reported, eclogite xenoliths sampled by kimberlites are relatively rare and are less abundant than peridotites and pyroxenites. Nonetheless, eclogites are the dominant xenoliths at five well-documented diamondiferous intrusives worldwide: Roberts Victor and Bobbejaan (South Africa), Zagadochnaya (Russia), Orapa (Botswana), and Koidu (Sierra Leone). Experimental work has demonstrated the transformation of basalt to eclogite sensu

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[1] Now at C. F. Mineral Research, Ltd., Kelowna, British Columbia, Canada.

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stricto at pressures and temperatures expected in the upper mantle [*Yoder and Tilley*, 1962; *Green and Ringwood*, 1967; *O'Hara and Yoder*, 1967]. This transformation enhances the importance of eclogite to recycling and chemical refurbishing of Earth's mantle, and yet the precise geochemical kinship between eclogite and the upper mantle remains controversial. Eclogite, or piclogite (olivine eclogite), has been proposed as a major component of the transition zone [*Anderson*, 1981]. Others [*Smyth and Caporuscio*, 1984; *Anderson*, 1989] have considered an eclogite cumulate layer in the upper mantle to be a possible product of a differentiated chondritic Earth and a source region for the majority of basaltic magma. The plausible link between eclogitic melts and the volatile large ion lithophile element and high field strength element signatures of kimberlites and lamproites is of primary importance [*Anderson*, 1984]. The wide range of major, minor, trace element, and isotopic compositions displayed by eclogites in the presence of a wide variety of peridotites supports the view for widespread mantle heterogeneity. An understanding of the chemical variations and the origins of eclogites given the widespread distribution of basalt in space and time is therefore essential in determining the origin, structure, composition, and evolution of silicate melts in Earth's interior.

There are at present at least six hypotheses proposed for the origin of mantle eclogites. The first five hypotheses are summarized by *Hills and Haggerty* [1989]: (1) metamorphic reaction from basalt to eclogite, either by transformation of gabbro previously underplated onto the lower crust or by subduction of ancient oceanic crust, (2) direct high-pressure crystallization, (3) exsolution of garnet from clinopyroxene, (4) decomposition of majorite, or (5) restite from the extraction of a high-Si component. The sixth hypothesis, proposed by *Anderson* [1984] and supported by the experimental work of *O'Hara and Yoder* [1967], suggests that eclogite is a possible source region for kimberlites, which may represent a late stage residual melt of an eclogitic cumulate layer.

Among these hypotheses the first two opposing models have provoked the most controversy. In the first model, mantle eclogites are interpreted by *Coleman et al.* [1965], *Helmstaedt et al.* [1972], *Smyth and Hatton* [1977], *Helmstaedt and Schulze* [1979], *Ater et al.* [1984], *Jagoutz et al.* [1984], *MacGregor and Manton* [1986], and *Shervais et al.* [1988] to represent the prograde metamorphic products of subducted Archean oceanic crust. This theory arises from the recognition that some eclogites and altered ocean floor basalts have broadly similar chemical compositions. Experimental studies on the synthesis of eclogites from high-Al basalt and gabbroic anorthosite by *Green* [1967] and *Green and Ringwood* [1967] bolster the metamorphic transformation hypothesis for eclogite xenoliths. In the second model, eclogites are interpreted to be products of igneous fractionation and are either high-pressure crystal cumulates, evolved liquids, or residues derived from mafic magmas that formed dikes within the upper mantle [*O'Hara and Yoder*, 1967; *Kushiro and Aoki*, 1968; *MacGregor and Carter*, 1970; *Hatton and Gurney*, 1977; *Smyth and Caporuscio*, 1984]. This hypothesis is based mostly on the interpretation by *MacGregor and Carter* [1970] that group I eclogites are igneous cumulates. Experimental evidence [*Yoder and Tilley*, 1962; *O'Hara and Yoder*, 1967] indicates that bimineralic garnet and clinopyroxene may represent the fractionation products of a liquid generated by the high-pressure partial melting of garnet lherzolites.

The compositional framework for eclogites from the Koidu Kimberlite Complex was established by *Tompkins and Haggerty* [1984] and *Hills and Haggerty* [1989]. In this contribution we focus on an expanded mineral compositional database and report on exsolution and complex zoning in garnet and clinopyroxene, partial melting and metasomatism, upper mantle silicate repositories for F, Cl, OH, Na, P, and Ti, a test of the mineral composition and modal recombination method to obtain bulk rock chemistries, and a refined treatment of depths of eclogite equilibration.

The Koidu Kimberlite Complex in eastern Sierra Leone is situated at 8°N, 11°W in the southwest corner of the Archean West Africa craton and is one of the four major cratonic nuclei hosting diamonds on the Africa continent. The Koidu Kimberlite Complex consists of three Mesozoic kimberlite pipes, a ring-dike structure, and multiple sets of en echelon kimberlite dikes that intruded into the Archean (~2.7 Ga) granitic basement of the Liberian Age Province [*Grantham and Allen*, 1960; *Morel*, 1979]. The kimberlite dikes strike N55°E and are both older and younger than the kimberlite pipes [*Tompkins and Haggerty*, 1984, and references therein]. The mantle xenoliths sampled by the Koidu kimberlite are unusual because of the apparent absence of peridotites and pyroxenites and the abundance of eclogites. Other xenoliths from Koidu include discrete garnet, pyroxene, ilmenite, and ilmenite-pyroxene intergrowths [*Tompkins and Haggerty*, 1984].

Analytical Techniques

Individual mineral compositions from 41 doubly polished thin sections were determined using the CAMECA SX-50 electron microprobe at the University of Massachusetts-Amherst. Standard microprobe calibration procedures were employed using a variety of natural and synthetic oxides and silicates. A qualitative energy-dispersive spectrometer (EDS) scan indicating the presence of detectable elements was conducted for all minerals prior to quantitative analyses. Kα radiation was used for each element (except for Sr, Zr, and Nb with which Lα radiation was utilized) using four wavelength-dispersive spectrometers and LiF, PC0, PET, and TAP analyzing crystals. The operating accelerating voltage was 15 kV, the takeoff angle was 40°, and the beam current (Faraday cup) was maintained at 15 nA with a minimum point beam of 2–5 μm in diameter. Counting intervals for each element generally ranged from 10 s for major elements to 50 s for minor elements. The PAP correction computation [*Pouchou and Pichoir*, 1984a, b] and a cation reduction scheme [*Deer et al.*, 1992] were used for all quantitative analyses. Water contents in hydrous minerals were determined by allocating H_2O molecules based on the quantity of other cations [*Deer et al.*, 1992]. All iron is reported as total FeO, except for amphibole, ilmenite, and acmite where Fe^{3+} corrections were made observing the uncertainties and restrictions as outlined by *Finger* [1972].

From the analyses obtained from the electron microprobe, end-members for garnet and clinopyroxene were calculated and presented in mole percent, following the formulation employed by *Hills and Haggerty* [1989]. Components for garnet analyses include mole percent of grossular (Gross, $Ca_3Al_2Si_3O_{12}$), pyrope (Pyr, $Mg_3Al_2Si_3O_{12}$), and almandine (Alm, $Fe_3Al_2Si_3O_{12}$): Gross = $100 \times (Ca/(Ca + Mg + Fe))$, Pyr = $100 \times (Mg/(Ca + Mg + Fe))$, and Alm = $100 \times (Fe/(Ca + Mg + Fe))$. End-member calculations for pyroxenes are presented in mole percent of wollastonite (Wo, $CaSiO_3$), enstatite (En, $MgSiO_3$), and ferrosilite (Fs, $FeSiO_3$): Wo = $100 \times$

(Ca/(Ca + Mg + Fe)), En = 100 × (Mg/(Ca + Mg + Fe)), and Fs = 100 × (Fe/(Ca + Mg + Fe)). Pyroxene data are also reported in terms of mole percent of jadeite (Jd, $NaAlSi_2O_6$), diopside (Di, $CaMgSi_2O_6$), and hypersthene (Hy, $MgFeSi_2O_6$): Jd = 100 × ((Na + ^{VI}Al + Cr)/(Na + ^{VI}Al + Cr + Ca + Mg + Fe + Mn)), Wo = 100 × (Ca/(Na + ^{VI}Al + Cr + Ca + Mg + Fe + Mn)), and Hy = 100 × ((Mg + Fe + Mn)/(Na + ^{VI}Al + Cr + Ca + Mg + Fe + Mn)).

Estimated whole rock compositions for the Koidu eclogites were computed using the modal recombination method. Because of the small size of the xenoliths and the large size of the constituent minerals, considerable effort was devoted to estimating the modal abundances of garnet and clinopyroxene of each sample by point counting in thin section and by determining the areas comprising these phases on hand specimen surfaces. The modal percent estimated was normalized based on the multiplication of the mole percent end-members and the corresponding densities. Individual mineral compositions and the normalized modal percent were then multiplied to give the reconstructed bulk compositions. The Ca-Tschermak component was excluded in the computation because most of the Koidu eclogitic pyroxenes contain <1 mol % Ca-Tschermak. Low-pressure mineral densities of end-members used in the modal recombination method are from *Robie et al.* [1966] and *Irifune et al.* [1986] and are as follows: grossular (density is 3.595 g/cm^3), pyrope (3.559 g/cm^3), almandine (4.318 g/cm^3), jadeite (3.315 g/cm^3), diopside (3.277 g/cm^3), enstatite (3.198 g/cm^3), and ferrosilite (3.992 g/cm^3).

Petrography

Among the ~2000 eclogites collected from Kimberlite Pipe 1, 41 samples were selected for detailed study (Table A1[1]). The eclogites are coarse-grained and equigranular and range from 2.5 × 2 × 2 cm to 6.5 × 4.5 × 1.5 cm in size. The eclogites consist mostly of garnet and clinopyroxene, but some have up to 15 modal % of secondary minerals. The garnet to pyroxene modal ratios range from 25:75 to 70:30. Most xenoliths are smooth-surfaced and ovoid in shape; others are blocky and subangular. The rounding and smoothing of the rock surfaces are attributed to magmatic corrosion and fluidization in the mantle [*Dawson*, 1980]. The small sizes and subangular morphologies of some eclogites reflect crushing and milling during treatment of the kimberlite to recover diamonds.

In hand specimen, garnets are subhedral to irregular in shape, average 4 mm in diameter, and vary from dark reddish brown to orange in color. Clinopyroxenes are mostly anhedral and irregular, average 3–4 mm in size, and are pale green to dark green in color. In some specimens, clear, emerald green to dark green pyroxenes are rimmed and veined by turbid, light green and fine-grained alteration intergrowths.

In thin section, garnet is colorless to light pink in color and shows no effect of partial melt. Most garnet is fractured, and some are veined by very fine grained amphibole, phlogopite, rutile, spinel, sulfide, and feldspar complexes of presumed secondary origin. Reaction along grain boundaries is reflected in the presence of kelyphite-like rims. One kelyphite type is dominated by an extremely fine grained, fibrous, opaque, and yellowish-brown material that is optically irresolvable. Another kelyphite type is mostly fine grained but is resolvable into high-birefringent aggregates of pyroxene, amphibole, phlogopite, spinel, and possibly plagioclase. Kelyphites are not restricted to garnet rims; some appear in reticulate patterns, and others fill veins throughout garnet grains. These textural characteristics are analogous to kelyphitized eclogitic garnets reported from Roberts Victor [*Kushiro and Aoki*, 1968; *MacGregor and Carter*, 1970], Kao [*Shervais et al.*, 1988], and Colorado-Wyoming [*Ater et al.*, 1984]. The kelyphites in Koidu may have resulted from metasomatism, retrograde reaction on decompression [*Mukhopadhyay*, 1991a], or reaction of garnet with infiltrating kimberlite.

Clinopyroxene examined in thin section is colorless to light green in color, slightly pleochroic, and typically more altered than coexisting garnets. In some eclogites the pyroxene has been severely altered to turbid, gray, or dull green mineral intergrowths and shows only occasional relics of pristine pyroxene. Most pyroxene has undergone various degrees of partial melting, reflected in the presence of spongy, symplectic textures. The extent of partial melting differs from rock to rock, with melt textures ranging from thin, discontinuous stringers in small concentrations to almost complete melting and reprecipitation of the original pyroxene. Pyroxene in nine eclogites displays thin, exsolved orthopyroxene lamellae which crosscut the entire grain. These lamellae are irregularly spaced and are parallel to apparent {100} or {010} directions.

In 18 eclogites (Table A1), very fine grained, tabular, oriented, brown to dark brown rutile lamellae are observed within the cores of garnet and clinopyroxene. The lamellae appear to be homogeneous and account for <2% by volume of individual host grains. The majority of rutile ranges from ~20–100 μm long in garnet to <50 μm long in clinopyroxene. Rutile is either acicular or is present as stubby, faceted prisms along apparent {111} or {110} garnet planes and {100} or {010} pyroxene planes. These textures are similar to rutile lamellae described in other eclogitic garnet and pyroxenes [*Griffin et al.*, 1971; *Meyer and Brookins*, 1971; *MacGregor*, 1979; *Tollo*, 1982; *Shervais et al.*, 1988; *Hills and Haggerty*, 1989] and to pyroxene lamellae in ultradeep majoritic garnets [*Haggerty and Sautter*, 1990]. Rutile lamellae have been interpreted by *Griffin et al.* [1971] as high-pressure exsolution products during the cooling of relatively Ti-rich silicates. Apatite lamellae have been also identified in host garnet and clinopyroxene and are described in the next section.

Well-cleaved, anhedral calcite is recognized in four eclogites (K86-36, K91-1, K91-5, and K91-46). Discrete sulfides (pyrite + pyrrhotite) are interstitial to silicate minerals. Minor amounts of amphibole, phlogopite, rutile, ilmenite, feldspar, and acmite are present in the eclogites and are metasomatic or late stage secondary minerals.

Most xenoliths in the eclogite suite are pristine, but some eclogites exhibit a very weak fracture fabric (Figure 1a). Garnets in these samples are generally unveined, and the pyroxenes are clear and well preserved. Minor alteration, if any, is restricted to grain peripheries where near-perfect triple junctions between garnet or pyroxene grains are observed. In contrast a strong fabric pervades 11 eclogites (Figure 1b). The

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[1] An electronic supplement of Tables A1–A19 may be obtained on a diskette or Anonymous FTP from KOSMOS.AGU.ORG. (LOGIN to AGU's FTP account using ANONYMOUS as the username and GUEST as the password. Go to the right directory by typing CD APEND. Type LS to see what files are available. Type GET and the name of the file to get it. Finally, type EXIT to leave the system.) (Paper 95JB01573, The petrography and mineral compositions of eclogites from the Koidu Kimberlite Complex, Sierra Leone, by A. T. Fung and S. E. Haggerty). Diskette may be ordered from American Geophysical Union, 2000 Florida Avenue, N.W., Washington, DC 20009; $15.00. Payment must accompany order.

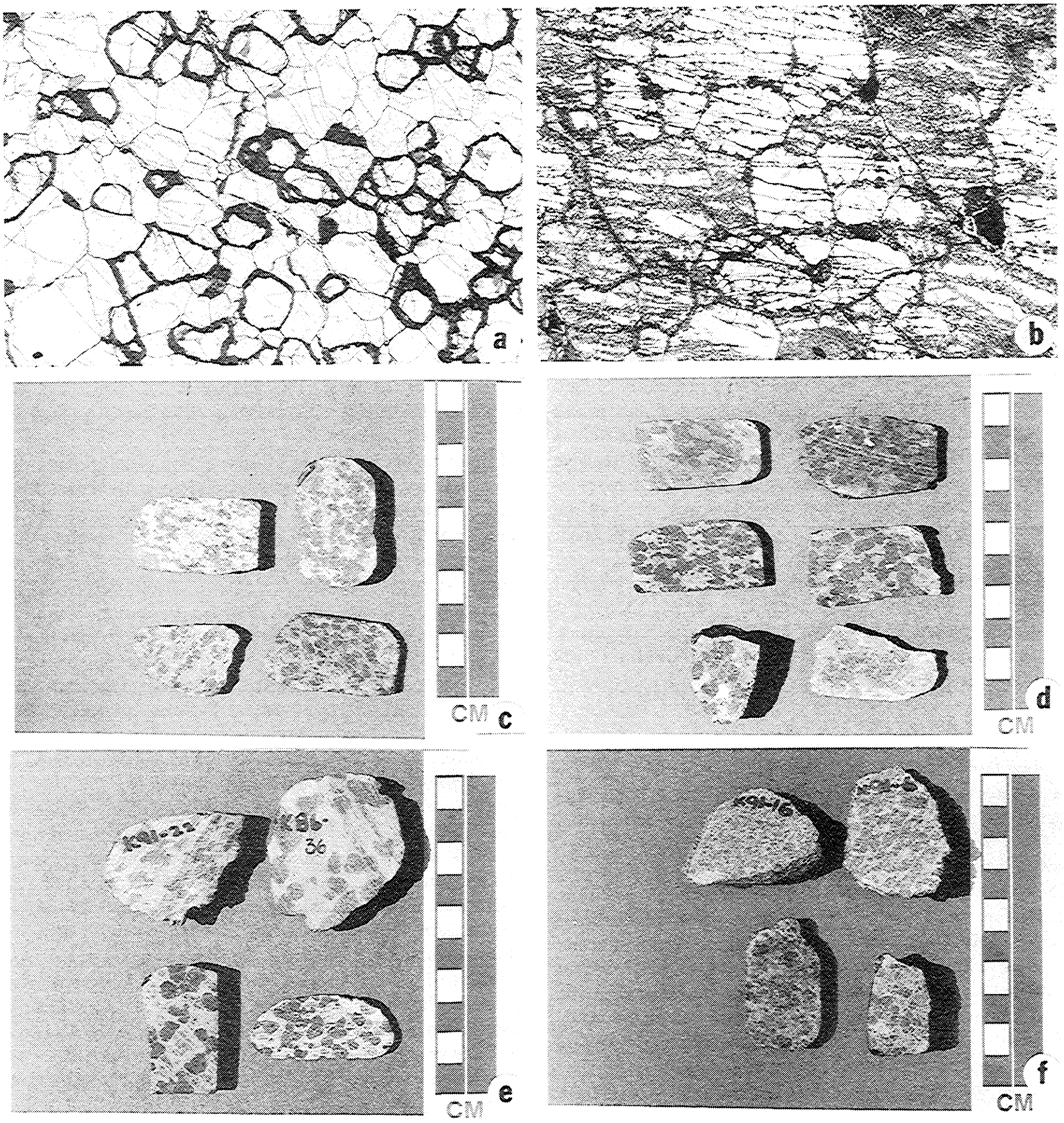


Figure 1. (a) Photomicrograph of Koidu eclogite K91-16 showing pristine and unaltered garnet and pyroxene with near-perfect triple junctions between grain boundaries. Transmitted plane-polarized light. Width of photomicrograph is 2 cm. (b) Photomicrograph of K91-36 showing a stong fabric that pervades garnet and pyroxene grains. Transmitted plane-polarized light. Width of photomicrograph is 2.4 cm. (c) The first group of Koidu eclogites characterized by highly altered clinopyroxene and tan or orange garnet. (d) A representative collection of the second group of eclogites represented by K91-58 and K91-59. (e) A representative collection of the third group of eclogites represented by K86-36 and K91-60. (f) A representative collection of the fourth group of eclogites represented by K86-60 and K91-33.

fabric crosscuts grain boundaries and consists of veined clusters of secondary amphibole, phlogopite, rutile, sulfide, spinel, calcite, and feldspar.

The mineral fabric classification proposed by *MacGregor and Carter* [1970] is not clearly applicable to the Koidu eclogites. *Hills and Haggerty* [1989] developed a hand specimen classification scheme based primarily on the presence or absence of accessory minerals: a bimineralic (garnet + clinopyroxene) suite which has no accessory minerals and is divided into three subgroups and seven suites of eclogites that have one or more accessory minerals (kyanite, coesite, corundum, amphibole, diamond, and graphite). The eclogites in this study are exclusively bimineralic and are divided into four groups, based on color, size, and relative proportions of garnet and pyroxene.

The first group of eclogites, represented by K91-1, is characterized by an overall light color and dull appearance (Figure

1c), with the garnet to pyroxene modal ratios of 33:67 to 55:45. Garnet is round and tan or orange in color and averages 4 mm in size. Pyroxene is light green and highly altered with occasional remnants of pristine pyroxene in the center. This group of xenoliths is similar to the group 1 eclogites described by *Hills and Haggerty* [1989]. The most distinguishing features of the second group of eclogites are the overall dark appearance, pristine garnet and pyroxene, and the blocky, angular shape of the xenoliths (Figure 1d); the shapes are probably imposed by milling of larger xenoliths. Fifteen samples in this study fall into this group; examples include eclogites K91-58 and K91-59. The grain boundaries between the dark red to tan garnets show near-perfect triple junctions. Pyroxene is very dark green and is slightly altered. Most xenoliths in this group have little or no fracture fabric. The eclogites are analogous to the group 2 eclogites recognized by *Hills and Haggerty* [1989]. Group 3 eclogites, represented by K86-36 and K91-60, differ from the second group in that the pyroxene is large (~1 cm), turbid, creamy and very pale green, and typically more abundant than garnet (Figure 1e); the average garnet to pyroxene modal ratio is 35:65. The xenoliths are typically smooth-surfaced and discoidal, and display abundant subparallel fractures crosscutting primary phases. Rounded orange garnet up to 1 cm in diameter is characteristic of the group 3 eclogites. The fourth group of eclogites, represented by K86-60 and K91-33, is dominated by pristine, subhedral, orange garnet and green pyroxene of different shades (Figure 1f). Garnet varies from 1 mm to 2 cm in size, and pyroxene averages 4 mm. Modal ratios of garnet to pyroxene are 35:65 to 70:30. A few xenoliths show the apple green color of clinopyroxene which is distinctive of the third group of eclogites described by *Hills and Haggerty* [1989].

Mineral Compositions

Garnet

Homogeneous garnet. The average chemical analyses of homogeneous garnets from the Koidu eclogites are given in Table A2. On the basis of the MgO contents of garnets the xenoliths are divided into two chemically distinct suites. The high-MgO eclogites (seven samples) contain pyropic garnets (16.5–20.2 wt % MgO) and have an average garnet composition of $Pyr_{65}Alm_{20}Gross_{15}$ (Figure 2a). Garnets in the low-MgO eclogites (27 samples) are almandine- and grossular-rich (5.3–13.2 wt % MgO) and vary from $Pyr_{20}Alm_{60}Gross_{20}$ to $Pyr_{35}Alm_{30}Gross_{35}$. Eclogite K91-13, which contains apatite lamellae in garnet, belongs to the low-MgO category with an average garnet composition of $Pyr_{39}Alm_{45}Gross_{16}$. This is consistent with the accessory mineral-bearing eclogite xenoliths described by *Hills and Haggerty* [1989], which are also characterized by low-MgO garnets. The high- and low-MgO garnets are separated by a significant compositional gap (Figure 2a), and the only eclogite that falls into this Mg gap is sample K91-18, which has a garnet composition of $Pyr_{53}Alm_{36}Gross_{11}$ (Figure 2b). Garnets in this rock contain 14.7 wt % MgO, 4.5 wt % CaO, and 17.6 wt % FeO (Table A2). Application of the statistical classification by *Dawson and Stephens* [1975] shows that six high-MgO garnets are identified as chrome-pyrope, 26 low-MgO garnets as calcic pyrope-almandine, and the remaining two fall in other categories. In terms of the hand specimen classification established above, all but one of the high-MgO eclogites belong to group 4, whereas those in groups 2 and 3 are exclusively of the low-MgO variety.

In general, high-MgO garnets have lower weight percent CaO (3.8–7.1) and FeO (9.9–14.0) than the low-MgO varieties, which contain 4.4–11.7 wt % CaO and 13.9–24.0 wt % FeO. The Koidu garnets analyzed in this study are very similar to those from *Hills and Haggerty* [1989] in terms of Ca, Mg, and Fe contents. High-MgO garnets in both studies fall exclusively within the field of group A eclogites as defined by *Coleman et al.* [1965]; the low-MgO garnets occupy the group B and C eclogite regions. This broad compositional range is typical of eclogitic garnets as shown in Figure 2c, which is a compilation of garnet data from different kimberlite localities: Roberts Victor [*Kushiro and Aoki*, 1968; *MacGregor and Manton*, 1986; *Caporuscio and Smyth*, 1990], Bellsbank [*Shervais et al.*, 1988; *Caporuscio and Smyth*, 1990], Jagersfontein [*Dawson and Smith*, 1986], Orapa [*Shee and Gurney*, 1979; *Tollo*, 1982], Yakutia [*Sobolev*, 1977; *Sobolev et al.*, 1968], Kansas [*Meyer and Brookins*, 1971], and Colorado-Wyoming [*Ater et al.*, 1984].

The compositions of crustal garnets in diamond-bearing and coesite-bearing eclogites from China [*Wang et al.*, 1990, 1992; *Shutong et al.*, 1992] are illustrated in Figure 2d. Also plotted for comparison are crustal metamorphic garnets associated with diamonds in a garnet-pyroxene rock and a pyroxene-carbonate rock from Russia [*Sobolev and Shatsky*, 1990] and two garnets in a coesite-bearing pyrope-quartzite from the western Alps [*Chopin*, 1984]. As is the case for upper mantle-derived garnets, these crustal garnets lie in all three of the *Coleman et al.* [1965] garnet regions. Attempts to classify eclogitic garnets into mantle and crustal varieties on the basis of pyrope contents are therefore unsuccessful. The overlap in compositions between upper mantle and crustal garnets suggests either that the kimberlitic eclogites in the group B and C fields (Figures 2c and 2d) are not of upper mantle origin or that the pyrope contents of upper mantle garnets are not confined to the lower limit of 55 mol % in the group A field, as proposed by *Coleman et al.* [1965]. Additional criteria are needed to distinguish between upper mantle and deep crustal garnets.

Zoned garnet. Among the 41 xenoliths in this study, six eclogites contain zoned garnets. The cores of these garnets are mantled by a second generation of garnet which is separated abruptly from the core by a thin moat of amphibole and phlogopite. Texturally, these mantled garnets resemble two zoned garnets described by *Hills and Haggerty* [1989]. Chemical analyses of the zoned garnets (Table A3) show that the cores are homogeneous but slight chemical variations are detected in the rims. Compared to homogeneous garnets, garnet cores from different xenoliths range from $Pyr_{26}Alm_{41}Gross_{33}$ to $Pyr_{45}Alm_{35}Gross_{20}$ and occupy the low-MgO field (Figure 3a). Corresponding garnet rims are more magnesian, with a substantial variation from $Pyr_{37}Alm_{38}Gross_{24}$ to $Pyr_{59}Alm_{27}Gross_{14}$. Rim garnets in three eclogites fall within the Mg gap; one rim garnet (sample KL91-37) is compositionally akin to high-MgO garnets.

Eclogite K91-5 has been studied in detail [*Fung*, 1994; *Haggerty et al.*, 1994] because of the presence of apatite lamellae in the cores of mantled garnets. Core garnets in this sample have a narrow compositional range ($Pyr_{30.1-30.7}Alm_{40.2-41.2}Gross_{28.2-29.3}$), but spectacular oscillatory zoning is displayed in garnet rims. Such zonal behavior has not previously been reported in either eclogitic or peridotitic minerals in kimberlitic xenoliths but is present in all six eclogites that contain zoned garnets. Because of the fine scale of the zones (micrometers to tens of micrometers wide), the garnet compositions in Table A3 are not absolute values for any one zone but are possibly averages of adjacent zones. As in the case of mantled garnets in other

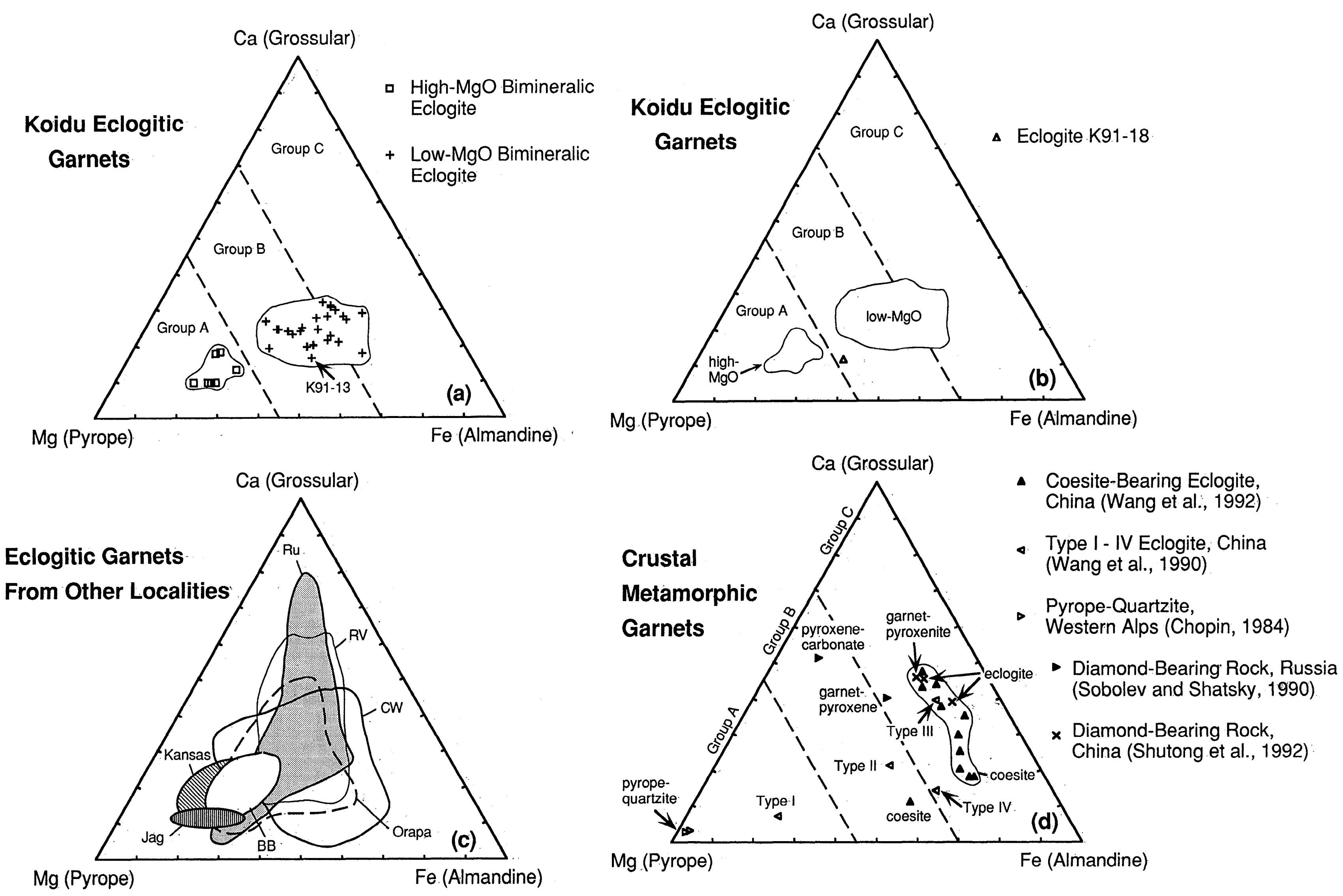


Figure 2. Garnet compositions plotted as mole percent pyrope-grossular-almandine. (a) Koidu eclogites. Sample K91-13 is apatite-bearing. Field boundaries for group A, B, and C eclogites are from *Coleman et al.* [1965]. (b) Koidu eclogite K91-18. Also plotted are the fields of high- and low-MgO Koidu eclogites. (c) Eclogites from various kimberlite localities: Ru, Russia; RV, Roberts Victor; CW, Colorado-Wyoming; Jag, Jagersfontein; BB, Bellsbank. (d) Crustal metamorphic rocks. See text for data sources and localities.

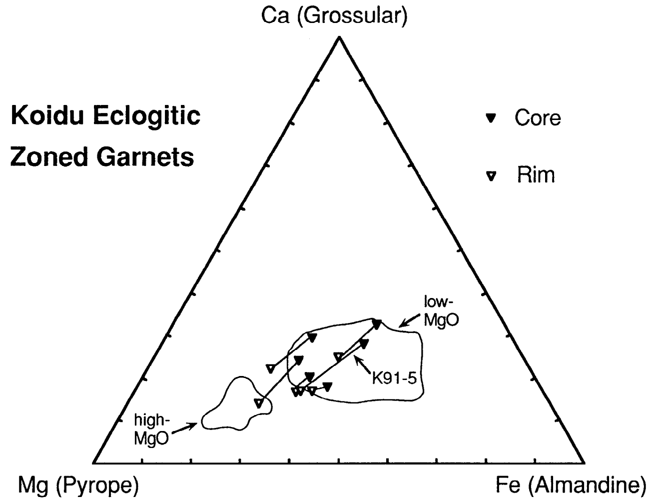


Figure 3a. Average compositions of the cores and rims of zoned garnets from Koidu eclogites plotted as mole percent pyrope-grossular-almandine. Also plotted are the fields of high- and low-MgO Koidu eclogites.

samples, the rim garnets in this eclogite are more pyropic ($Pyr_{43-56}Alm_{29-38}Gross_{15-20}$) than core garnets.

Major element profiles from the edge of the core of a zoned garnet in eclogite K91-5 across the moat assemblage and into an adjacent pyroxene show that FeO, CaO, and MgO are relatively constant in the core of the garnet (Figure 3b). The moat is enriched in MgO and depleted in CaO and FeO, which is consistent with the presence of phlogopite and amphibole. In the broadest zone of the rim garnet there are marked increases in FeO and CaO and complementary decreases in MgO (arrow 1 on Figure 3b). Other zones, however, exhibit a different zoning pattern with MgO increasing at the expense of FeO and CaO (arrows 2 and 2′ on Figure 3b). Both types of zonal behavior are present in an intermediate zone between 2 and 2′. In detail, FeO is the most sensitive to oscillatory zonation, whereas MgO and CaO have a longer wavelength variation than FeO across the entire profile. The contact zone between garnet and adjacent pyroxene is compounded by partial melting of the pyroxene and has a highly variable composition, but the profiles within unaffected pyroxene cores are flat. Profiles of SiO_2 and Al_2O_3 at the same distance are flat for both core and rim garnets. At the moat assemblage and the garnet-pyroxene contact zone, increases in SiO_2 are generally matched by decreases in Al_2O_3. This complementary pattern is consistent with the presence of amphibole, phlogopite, and pyroxene. *Haggerty et al.* [1994] interpreted the moat as the product of an upward metasomatic front, which is more volatile than the main body of melt transporting these fluids. The garnet overgrowths on the moat probably precipitated when the main part of the melt passed through the eclogite body.

Na, Ti, and P in garnet. The distribution and interrelationships of Na_2O, TiO_2, and P_2O_5 in garnets have been previously explored [*Ringwood and Major*, 1971; *Sobolev and Lavrent'ev*, 1971; *Thompson*, 1975; *Bishop et al.*, 1976, 1978; *Reid et al.*, 1976; *Haggerty et al.*, 1994] in an attempt to identify substitution schemes in upper mantle garnets. Na_2O concentrations (0.02–0.17 wt %) in Koidu garnets are similar to those of eclogitic garnets studied by *Sobolev and Lavrent'ev* [1971], *Reid et al.* [1976], *Bishop et al.* [1976, 1978], *Tollo* [1982], and *Hills and Haggerty* [1989]; the rim garnet of eclogite K91-26 is an exception which has a high Na_2O content of 0.33 wt %. *Sobolev* [1977] recorded up to 0.75 wt % TiO_2 in an eclogitic garnet from Udachnaya (Russia), but garnets from Koidu have lower weight percent TiO_2 (0.04–0.39). The P_2O_5 contents of these garnets vary from 0.02 to 0.26 wt %, except for sample K91-38, which contains 0.38 wt % P_2O_5 and is more phosphorus-rich than any eclogitic garnet documented in the literature.

The weight percent Na_2O of Koidu garnets increases with weight percent TiO_2, and low-MgO garnets are generally more enriched in both elements than high-MgO garnets (Figure 4a). Except for eclogite K91-7, the rims of all zoned garnets are more Na_2O- and TiO_2-rich than corresponding cores. The Na_2O- and TiO_2-rich garnet rims imply that if the melt infiltration model for mantled garnets is correct, the main body of the melt which precipitated the magnesian rims was probably also enriched in Na_2O and TiO_2. The strong positive correlation in a plot of weight percent Na_2O versus weight percent (TiO_2 + P_2O_5) in Figure 4b further indicates that Na in garnet may indeed be explained by coupled substitution with Ti and P, as proposed by *Bishop et al.* [1978]. With respect to these minor elements, garnets in diamondiferous [*Reid et al.*, 1976] and diamond-free eclogites [*Bishop et al.*, 1976] from southern Africa (Figures 4c and 4d) are similar to the Koidu garnets. In

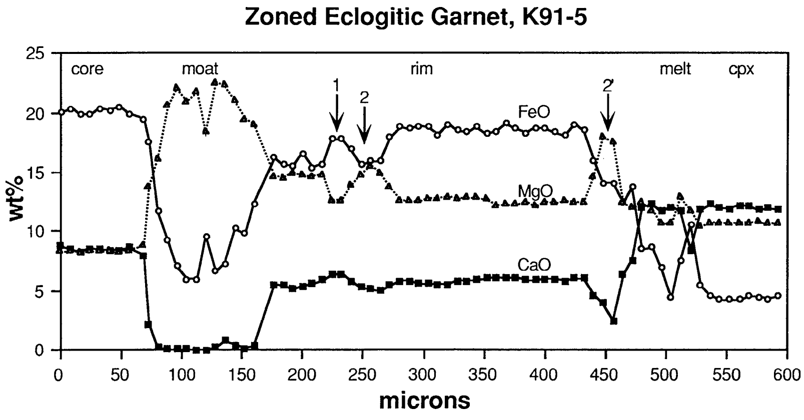


Figure 3b. Quantitative variations in weight percent FeO, MgO, and CaO from the core of a garnet across a moat assemblage and into an adjacent pyroxene from eclogite K91-5. Substitution behavior (arrows 1, 2, and 2′) among these elements is discussed in the text.

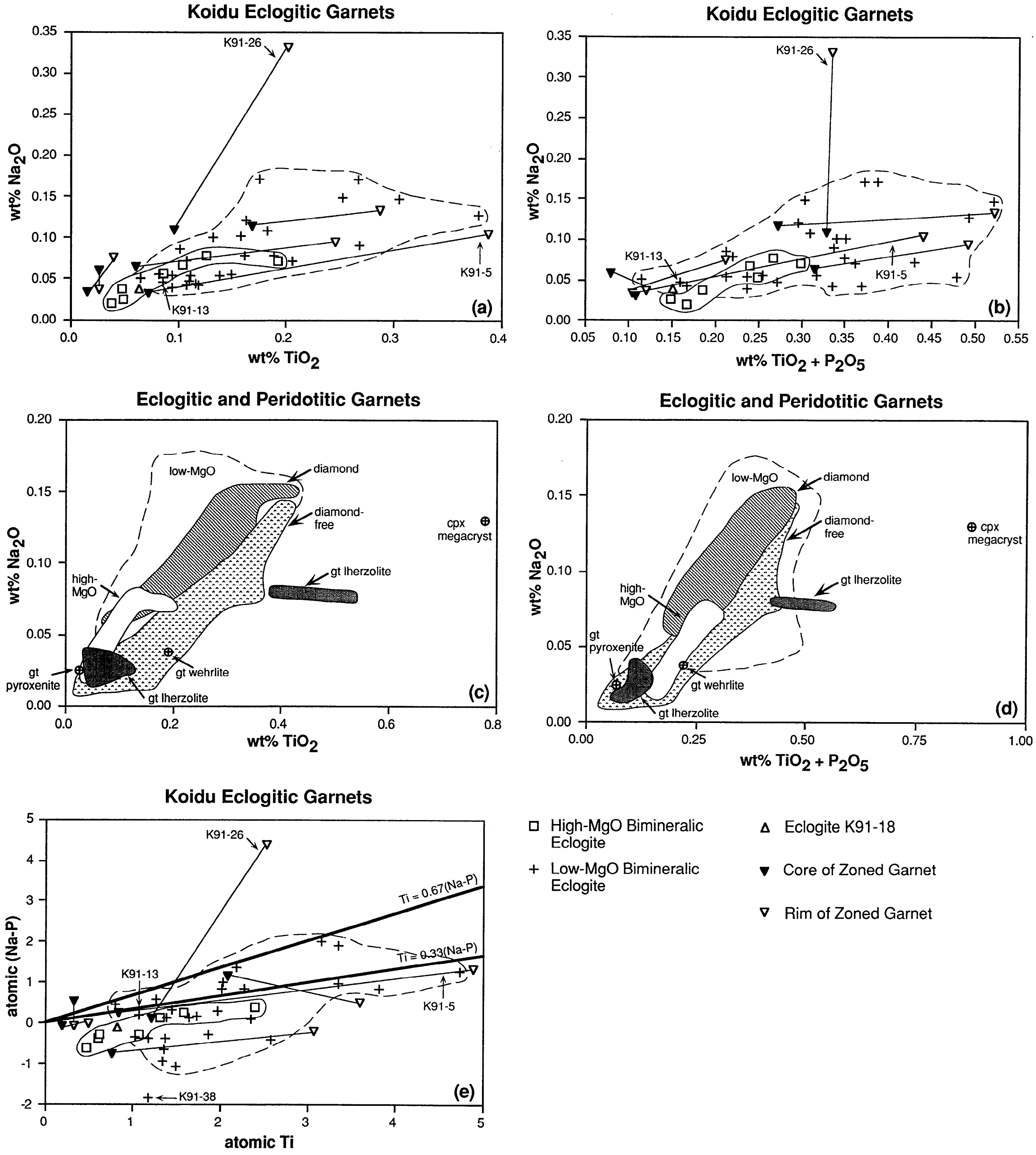


Figure 4. (a) Plots of weight percent TiO_2 versus Na_2O and (b) weight percent (TiO_2 + P_2O_5) versus Na_2O for Koidu eclogitic garnets. Samples K91-5 and K91-13 are apatite-bearing. The rim garnet of K91-26 is a prominent outlier. (c) Plots of weight percent TiO_2 versus Na_2O and (d) weight percent (TiO_2 + P_2O_5) versus Na_2O showing the fields of eclogitic and peridotitic garnets. See text for data sources and localities. (e) Plot of atomic Ti versus (Na-P) for Koidu eclogitic garnets. K91-38 and the rim garnet of K91-26 are outliers.

addition, garnets in a garnet pyroxenite, a garnet wehrlite, and the majority of garnet lherzolites from southern Africa [*Bishop et al.*, 1976] fall close to the linear eclogite trend. The only two outliers are a garnet lherzolite and a clinopyroxene megacryst [*Bishop et al.*, 1976] which contain higher TiO_2 and P_2O_5. Unlike igneous garnets in *Thompson*'s [1975] study, correlations do not exist between Na_2O and P_2O_5 for the Koidu garnets, nor is there a relationship in the diamond-bearing eclogites studied by *Reid et al.* [1976]. Nonetheless, the trace abundances of Na, Ti, and P in these mantle garnets confirm the suggestion by *Reid et al.* [1976] that substitution of these elements in the garnet structure is feasible at high pressure and that garnet may be a storage site for Na, Ti, and P in upper mantle rocks.

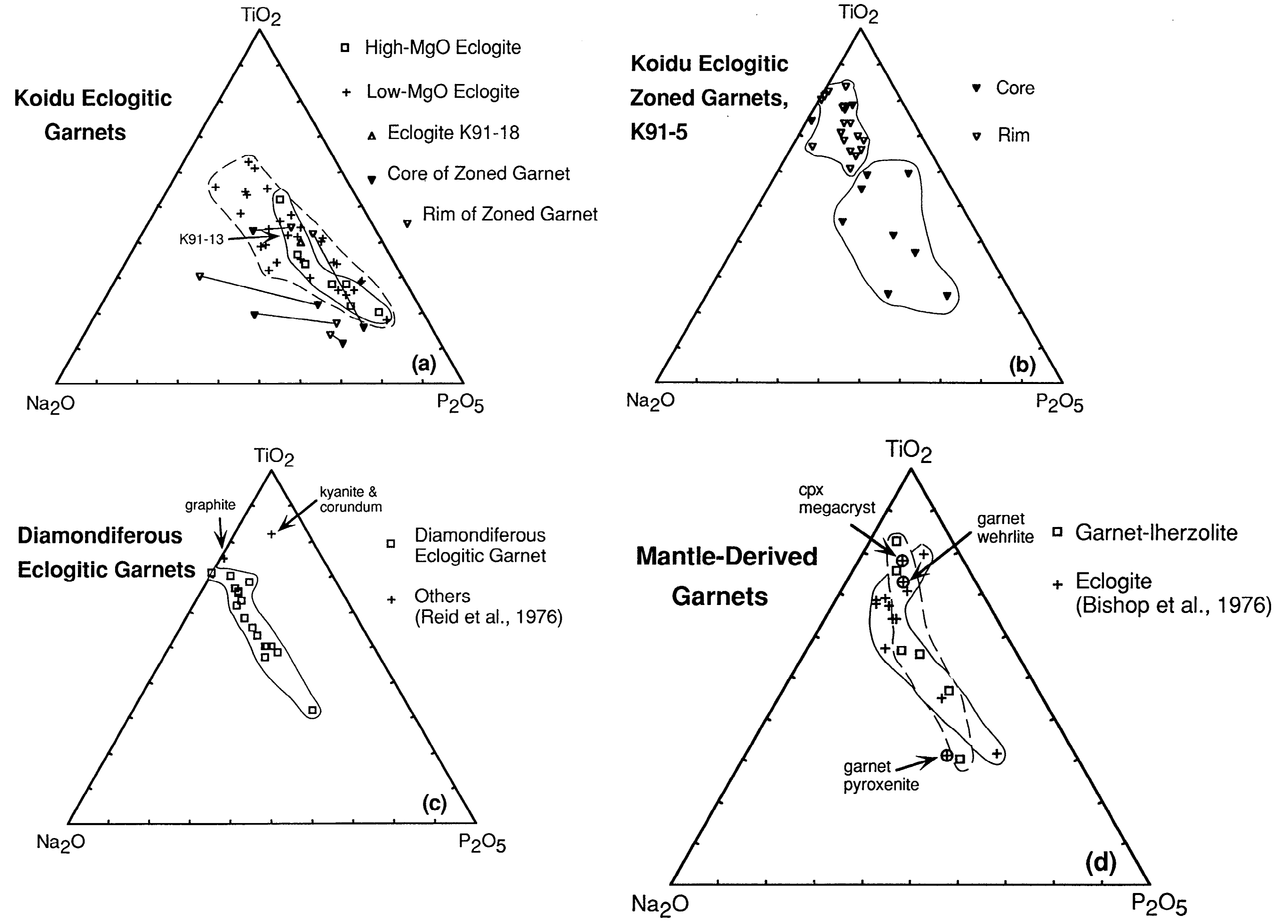


Figure 5. Ternary plot of weight percent TiO_2-Na_2O-P_2O_5. (a) Koidu eclogitic garnets. Sample K91-13 is apatite-bearing. (b) Cores and rims of zoned garnets from Koidu eclogite K91-5. (c) Garnets from diamondiferous eclogites, a graphite-bearing eclogite, and a kyanite- and corundum-bearing eclogite. All data are from southern Africa [*Reid et al.*, 1976]. (d) Eclogitic and lherzolitic garnets from southern Africa [*Bishop et al.*, 1976].

Relationships of the Koidu garnets in a plot of atomic Ti versus atomic (Na-P) differ from those noted by *Bishop et al.* [1978] in eclogites from South Africa (Figure 4e). *Bishop et al.* [1978] showed that almost all of the analyzed garnets fall near the line $Ti = 0.67(Na-P)$ and concluded that a common factor controls the distribution of these elements. Data for Koidu garnets lie closer to a line of slope 0.33, comparable to some of the peridotitic garnets studied by *Bishop et al.* [1978]. Some of the low-MgO garnets, however, scatter below the $Ti = 0.33(Na-P)$ line. Because atomic Ti is greater than atomic (Na-P), *Bishop et al.* [1978] postulated that substitutions of $Ti^{4+} \rightarrow Si^{4+}$, or $Ti^{3+} \rightarrow Al^{3+}$ may be invoked. Although coupled substitutions of the form $({}^{VIII}Ca + {}^{IV}Si) \rightarrow ({}^{VIII}Na + {}^{IV}P)$ and $({}^{VIII}Ca + {}^{VI}Al) \rightarrow ({}^{VIII}Na + {}^{VI}Ti)$ have been proposed by numerous workers [*Thompson*, 1975; *Bishop et al.*, 1976, 1978; *Reid et al.*, 1976], there is no apparent trend among the Koidu garnets in plots of (Ca + Si) versus (Na + P) and (Ca + Al) versus (Na + Ti). Significant correlations for other possible substitution schemes involving both major and minor elements (e.g., $(2Si + Fe) \rightarrow (Na + Ti + P)$) have also not been observed.

In a ternary diagram of weight percent TiO_2-Na_2O-P_2O_5, homogeneous garnets from Koidu show a linear trend from the Na_2O-TiO_2 sideline toward the P_2O_5 apex (Figure 5a). The fields for high- and low-MgO garnets overlap, with some low-MgO garnets more depleted in the P_2O_5 component. With the exception of eclogites K86-36 and K91-7, cores of zoned garnets are enriched in P_2O_5 relative to rim garnets. The same linear pattern is displayed by zoned garnets in eclogite K91-5 (apatite-bearing) which have P_2O_5-rich cores and TiO_2-rich rims (Figure 5b). The average P_2O_5 component of rim garnets in eclogite K91-5 is the lowest of all garnets in this eclogite suite.

Eclogitic and lherzolitic garnets from southern Africa (Figures 4c and 4d) show a similar compositional behavior to garnets from Koidu. In a ternary plot of weight percent TiO_2-Na_2O-P_2O_5 (Figure 5c) a graphite-bearing eclogite and 17 diamondiferous eclogitic garnets [*Reid et al.*, 1976] lie in a well-defined trend from the Na_2O-TiO_2 sideline toward the P_2O_5 apex. Garnet from a kyanite- and corundum-bearing eclogite [*Reid et al.*, 1976] falls outside the trend but is similar to rim garnets in eclogite K91-5 in terms of the relative proportions of these components. The linear pattern is maintained for other eclogitic and lherzolitic garnets [*Bishop et al.*, 1976] as illustrated in Figure 5d. Also plotted in Figure 5d are garnets from a clinopyroxene megacryst, a garnet wehrlite, and a garnet pyroxenite [*Bishop et al.*, 1976] which lie close to the eclogite and lherzolite trends.

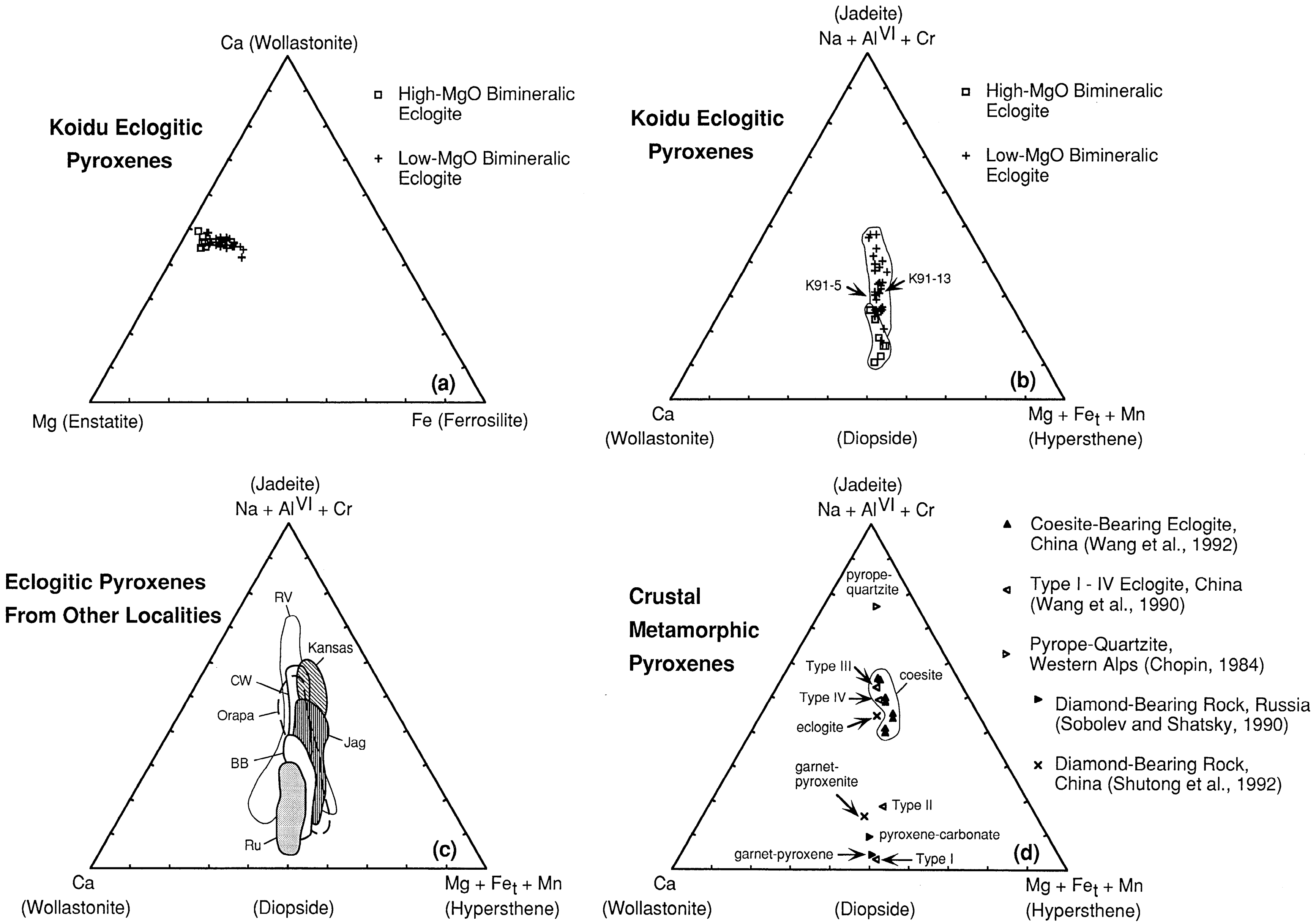


Figure 6. Pyroxene compositions from Koidu eclogites plotted (a) as mole percent wollastonite-enstatite-ferrosilite and (b) as mole percent jadeite-wollastonite-hypersthene. (c) Eclogitic pyroxenes from various kimberlite localities. Ru, Russia; RV, Roberts Victor; CW, Colorado-Wyoming; Jag, Jagersfontein; BB, Bellsbank. (d) Pyroxenes from crustal metamorphic rocks. The sources and localities are the same as in Figure 2d.

Clinopyroxene

Homogeneous clinopyroxene. Average chemical analyses of homogeneous clinopyroxenes are listed in Table A4. Although there is a wide range of weight percent CaO (12.6–22.0), MgO (8.63–15.8), and FeO (1.38–8.13), pyroxenes in the high- and low-MgO eclogites do not differ significantly ($Wo_{42}En_{41}Fs_{17}$ to $Wo_{49}En_{48}Fs_3$) in terms of the Ca-Mg-Fe component (Figure 6a). However, nonquadrilateral variations in weight percent Na_2O (1.64–7.18) and Al_2O_3 (2.06–12.42) are significant with considerable substitution of jadeite in diopside. Pyroxenes in the high-MgO eclogites are diopsidic ($Jd_{11}Di_{89}$–$Jd_{26}Di_{74}$), whereas those in the low-MgO eclogites ($Jd_{20}Di_{80}$–$Jd_{48}Di_{52}$) vary from jadeitic-diopside to omphacite (Figure 6b). Such a wide compositional range has been reported in other Koidu eclogitic pyroxenes [*Hills and Haggerty*, 1989]. On the basis of the *Stephens and Dawson* [1977] statistical classification, three pyroxenes from high-MgO eclogites are classified as chrome-diopside, two are classified as diopside, and two may be designated as either diopside or chrome-diopside; 16 low-MgO pyroxenes are classified as jadeitic diopside, 12 are classified as omphacite, and the remaining two fall in other categories.

Clinopyroxene analyses from kimberlitic eclogites are not as comprehensive as garnet analyses in the literature. Nonetheless, a compilation of pyroxene data from Roberts Victor [*Kushiro and Aoki*, 1968; *MacGregor and Manton*, 1986], Bellsbank [*Shervais et al.*, 1988], Yakutia [*Sobolev*, 1977], Jagersfontein [*Dawson and Smith*, 1986], Orapa [*Shee and Gurney*, 1979; *Tollo*, 1982], Kansas [*Meyer and Brookins*, 1971], and Colorado-Wyoming [*Ater et al.*, 1984] are shown in Figure 6c. In this jadeite-wollastonite-hypersthene ternary, most eclogitic pyroxenes lie on the jadeite-diopside join, similar in behavior to those from Koidu. Pyroxenes, coexisting with crustal diamond or coesite, do not differ in major element compositions from upper mantle-derived pyroxenes (Figure 6d).

Partially melted clinopyroxene. Most clinopyroxenes in this suite of eclogites have undergone various degrees of partial melting, which is reflected in the presence of spongy, symplectic textures. Compositions of partially melted pyroxenes (Table A5) range from Jd_8Di_{92} to $Jd_{30}Di_{70}$ with no differences between high- and low-MgO eclogites. The partially melted pyroxenes are almost identical to primary pyroxenes in terms of the Ca-Mg-Fe component but are distinctly more diopsidic (Figure 7a) than unmelted counterparts (Figure 6b). This is consistent with primary clinopyroxenes in diamondiferous eclogites [*Reid et al.*, 1976] which are jadeitic relative to other pyroxenes that have undergone partial melting. In eclogite K91-5 (apatite-bearing) a fine-grained, intergranular assemblage is present in the partial melt textures and has been determined to be sanidine and an apparent Na-Al glass (Table A6). A similar assemblage of albite and sanidine has been identified in partial melts in eclogite K91-18; albite has also been identified in the melt textures in K91-64.

In addition to the intermediate garnet composition between high- and low-MgO garnets, eclogite K91-18 is unusual because of the distribution of partial melt textures in pyroxene. Figure 7b is a schematic drawing of a thin section of this melt rock, showing that only a portion of the eclogite contains partially melted pyroxene; pyroxene in the rest of the xenolith is pristine and homogeneous. The partially melted part of the rock may represent a melt front, a fault contact, a vein, or a dike or may be the product of volatile-induced melting. Compositions of unaltered pyroxene are almost identical to the rim composition of zoned pyroxene.

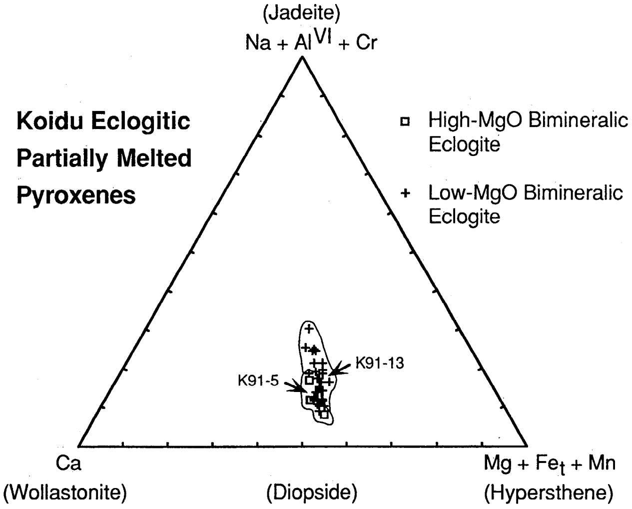


Figure 7a. Compositions of partially melted pyroxenes from Koidu eclogites plotted as mole percent jadeite-wollastonite-hypersthene. Samples K91-5 and K91-13 are apatite-bearing.

Zoned clinopyroxene. Average chemical analyses of the cores and rims of zoned pyroxene in four eclogites are listed in Table A7. In this section, distinct symplectite-like, partial melt textures are present in the cores of pyroxenes, which are surrounded by an overgrowth of recrystallized rims (Figure 8a). Ca, Mg, and Fe components for core and rim pyroxenes are almost indistinguishable. Variations in compositions are mainly in the jadeite and diopside components, with the cores of different xenoliths ranging from $Jd_{10}Di_{90}$ to $Jd_{19}Di_{81}$ and the rims ranging from $Jd_{17}Di_{83}$ to $Jd_{32}Di_{68}$ (Figure 8b).

Major element profiles of weight percent (CaO + MgO) and weight percent (Na_2O + Al_2O_3) across a zoned pyroxene grain in eclogite K91-37 show that chemical concentrations in pyroxene rims are fairly constant, with slight increases in CaO and MgO matched by slight decreases in Na_2O and Al_2O_3 (arrows 1 and 1′ on Figure 8c). The zonal behavior in the partially melted core, however, is complex and oscillatory. Compared with rim pyroxenes, the core pyroxene is CaO- and MgO-rich and Na_2O- and Al_2O_3-poor. Exceptions are two data points on the profiles which display lower CaO and MgO abundances than rim pyroxenes (arrows 2 and 2′ on Figure 8c). One of these analyses also contains higher Na_2O and Al_2O_3 contents than the pristine rims (arrow 2 on Figure 8c). The contacts between the core and rim in zoned pyroxene are abrupt, resembling the core-rim boundary in zoned garnets. Coarse skeletal pyroxene crystals, such as those in eclogite K91-18, imply rapid recrystallization following the melt event, which we interpret to be transient.

Garnet-Clinopyroxene Pairs

The compositions of garnet and pyroxene from eclogites in this study and those by *Hills and Haggerty* [1989] are expressed in a plot of mole percent pyrope versus mole percent jadeite (Figure 9a). The high- and low-MgO eclogites in both studies are separated by a prominent gap which reinforces the bulk chemical Mg classification proposed by *Hills and Haggerty* [1989]. Three amphibole- and phlogopite-rich low-MgO eclogites contain <15 mol % jadeite and fall outside the low-

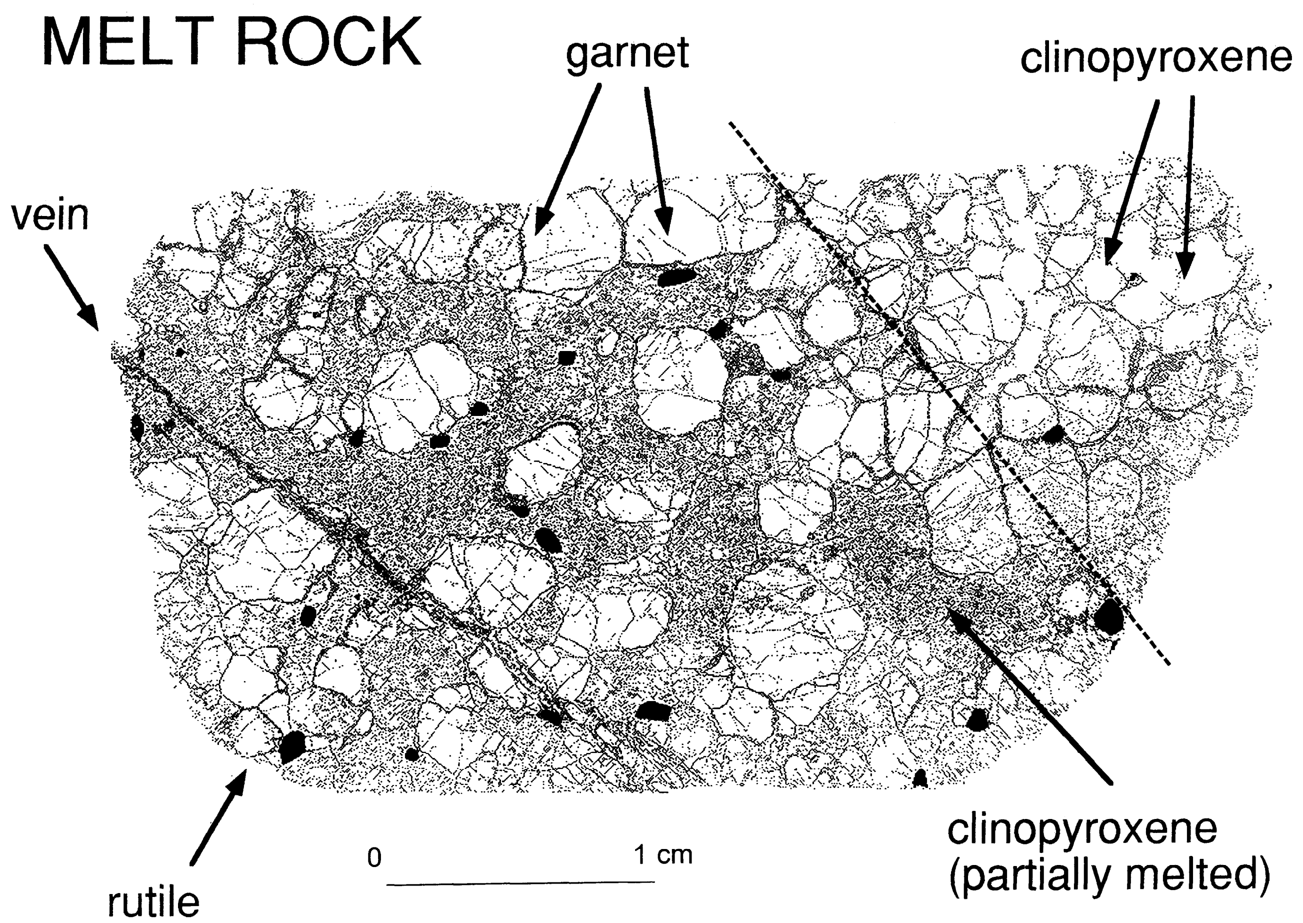


Figure 7b. A schematic drawing of a thin section of the melt rock, eclogite K91-18. Pyroxenes to the left of the dashed line display partial melt textures; pyroxenes to the right are pristine and homogeneous.

MgO cluster. Diamondiferous and kyanite-bearing varieties also form two separate clusters. Other outliers in the diagram include a kyanite + coesite eclogite (KEC86-QB1), a corundum + kyanite eclogite (KEC86-CB-6B), and a granulite (KGR86-76) which are described by *Hills and Haggerty* [1989].

The pyrope and jadeite components of zoned garnet and pyroxene in the Koidu eclogites are illustrated in Figure 9b, along with the boundaries of the eclogite fields established (Figure 9a). Compositions of the melt rock (K91-18) and three other eclogites containing zoned minerals plot exclusively in the Mg gap; others lie mostly in the low-MgO field. The wide compositional range of zoned garnet and clinopyroxene shown in this diagram is consistent with observations made in previous sections.

In terms of the same pyrope-jadeite relationship, diamondiferous eclogites from southern Africa [*Reid et al.*, 1976], Orapa [*Robinson et al.*, 1984], and Yakutia [*Jerde et al.*, 1993] occupy a region close to the Koidu diamondiferous eclogites (Figure 9c). Also plotted for comparison are crustal eclogites previously shown in Figures 2d and 6d, which lie close to the Koidu eclogite fields. The only significant outliers are coesite-bearing pyrope-quartzites from the western Alps [*Chopin*, 1984] which average 95 mol % pyrope and 75 mol % jadeite. As noted in previous garnet and pyroxene ternary diagrams, it is unclear why mantle and crustal eclogites do not differ significantly in mineral compositions.

Garnets in most diamond-bearing peridotitic mantle rocks are purple or purple-red with high weight percent MgO (20–26), high weight percent Cr_2O_3 (3–16) and low weight percent CaO (0–4) and are classified as G10 garnets [*Dawson and Stephens*, 1975]. Diamondiferous eclogites from Roberts Victor, on the other hand, are characterized by garnets with $\geq$0.09 wt % Na_2O and pyroxenes with $\geq$0.08 wt % K_2O, as recog-

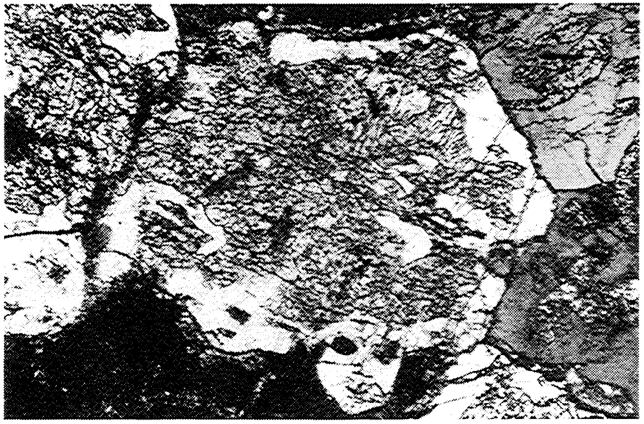


Figure 8a. Photomicrograph of a zoned pyroxene in eclogite K91-18, showing the partially melted core overgrown by a pristine and homogeneous rim. Transmitted crossed polars. Width of photomicrograph is 4.44 mm.

nized by *McCandless and Gurney* [1989]. A compilation of garnet and pyroxene data shows that such Na_2O and K_2O enrichment is also typical of diamondiferous eclogites from other localities (Figure 10a). Likewise, the weight percent TiO_2 ($\geq$0.3) in most diamond-bearing eclogites [*Reid et al.*, 1976; *Robinson et al.*, 1984; *Hills and Haggerty*, 1989; *Jerde et al.*, 1993] is more enriched than in the diamond-free variety (Figure 10b). Because of Na, Ti, and P substitution in garnet, these elements may be important high-pressure indicators associated with diamond genesis. It is also noted that garnets in diamondiferous eclogites contain higher Si (>2.95) and lower ^{IV}Al cations (<0.06) than diamond-free eclogites (Figure 10c). The compositions of eclogitic garnet and clinopyroxene may therefore be used to evaluate the diamond potential of a kimberlite and may aid in prospecting and exploration activities.

Apatite

Apatite crystals have been identified within the cores of garnet and clinopyroxene in eclogite K91-5 and are discussed by *Fung and Haggerty* [1993] and *Haggerty et al.* [1994]. Apatite in three additional eclogites (KEC-81-12, K91-13, and K91-64) have now been identified: KEC-81-12 contains apatite lamellae in garnet and pyroxene, K91-13 contains apatite lamellae in garnet, and K91-64 contains apatite lamellae in partially melted pyroxene. The petrography and mineral compositions of garnet and pyroxene in KEC-81-12 have been determined by *Hills* [1988]. This rock is not included in the eclogite suite (41 samples) in this study but is incorporated into discussions here because of the presence of apatite.

Examined in thin section, apatite is colorless and accounts for <2% by volume in individual host garnet and pyroxene grains. Apatite occurs either as rods (~200 × 20 μm) or faceted prisms (5–20 μm) along apparent {111} or {110} garnet planes and {100} or {010} pyroxene planes. Eclogite KEC-81-12 is unusual in that both apatite and rutile lamellae are present in garnet and pyroxene (Figure 11). The oriented textures exhibited by these apatites are suggestive of exsolution and are similar to pyroxene lamellae in ultradeep, majoritic garnets [*Haggerty and Sautter*, 1990] and to rutile lamellae in eclogitic garnet and pyroxene [e.g., *Tollo*, 1982; *Shervais et al.*, 1988; *Hills and Haggerty*, 1989]. Similar to K91-5, apatites in KEC-81-12, K91-13, and K91-64 differ from those in igneous garnets which are randomly distributed and may contain melt and/or mineral inclusions [*Irving and Frey*, 1978; *Embey-Isztin et al.*, 1985].

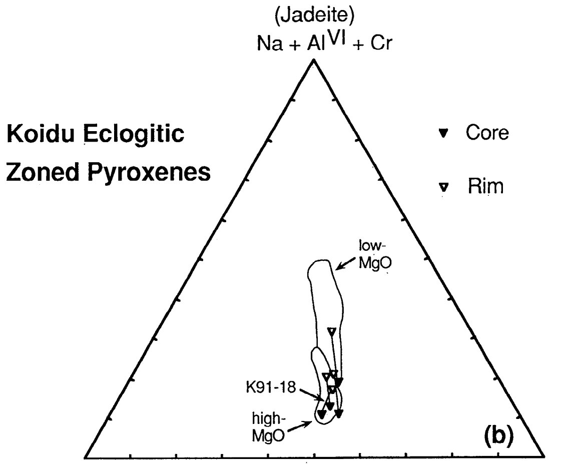


Figure 8b. Average compositions of the cores and rims of zoned pyroxenes from Koidu eclogites plotted as mole percent jadeite-wollastonite-hypersthene. Also plotted are the fields of high- and low-MgO eclogites. Sample K91-18 is the melt rock.

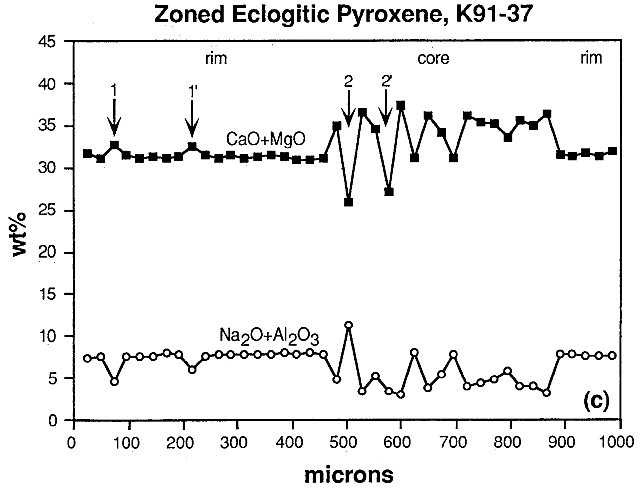


Figure 8c. Quantitative variations in weight percent (CaO + MgO) and (Na_2O + Al_2O_3) across a zoned pyroxene from eclogite K91-37. Substitution behavior (arrows 1, 1′, 2, and 2′) among these elements are discussed in the text.

Representative chemical analyses of apatites in K91-5, K91-64, and KEC-81-12 are given in Table A8. Data for eclogite K91-13 are not available because the apatites exposed at the surface are too small for quantitative microprobe analyses; high Ca and P in the lamellae have been confirmed by EDS. Apatite in the three analyzed specimens is homogeneous and differs slightly in weight percent P_2O_5 (39.4–41.9) and CaO (52.4–58.5). On the basis of F, Cl, and calculated H_2O concentrations, apatite in K91-5 (1.72 wt % F, 0.80 wt % H_2O) and KEC-81-12 (1.80 wt % F, 0.78 wt % H_2O) is classified as hydroxy-fluor-apatite, and apatite in K91-64 (1.73 wt % H_2O) is classified as hydroxy-apatite [*Deer et al.*, 1992]. Apatite contains <0.05–0.55 wt % SrO, which is much lower than SrO contents in apatite in metasomatic assemblages in basanites [*O'Reilly and Griffin*, 1988] and kimberlites [*Exley and Smith*, 1982]. Unlike some metasomatic apatite (F/Cl < 1) in basanites [*O'Reilly and Griffin*, 1988], the F/Cl ratios of samples K91-5 and KEC-81-12 are >1 and are analogous to other upper mantle, melt-derived megacrystic apatite [*Baker and Wyllie*, 1992]. On the basis of the crystallographic control of apatite, the absence of melt or crystal inclusions, F/Cl ratios, and distinctly low Sr concentrations, *Haggerty et al.* [1994] conclude that the Koidu eclogitic apatite is upper mantle-derived and may have originated from garnet and pyroxene by exsolution. The garnet-like structure in $Na_3Al_2P_3O_{12}$, first proposed by *Thompson* [1975], suggests that P may be substituted into garnet at mantle pressures. Substitution schemes involving P in pyroxene require exploration, but the presence of apatite-bearing pyroxene suggests that it may also be feasible to incorporate phosphorus into the pyroxene structure. This is in fact demonstrated by phosphoran pyroxene in the highly reduced basalts of Disko Island, Greenland [*Goodrich*, 1984]. Apatite exsolution in the Koidu eclogitic minerals is most likely to be related to decompression and cooling [*Haggerty et al.*, 1994].

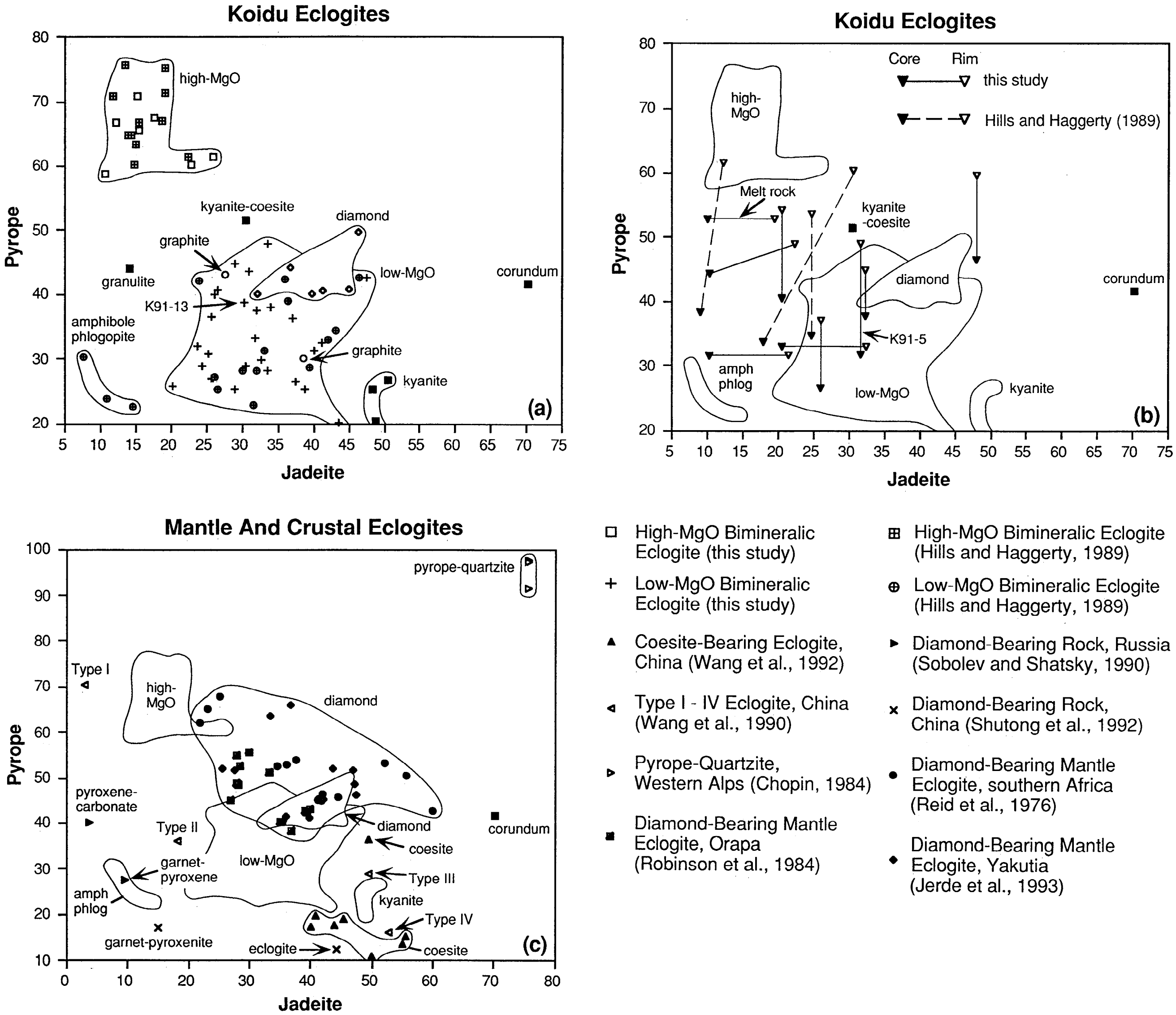


Figure 9. Plot of mole percent jadeite versus mole percent pyrope. (a) Koidu eclogites. Sample K91-13 is apatite-bearing. Accessory mineral-bearing eclogites are from *Hills and Haggerty* [1989]. (b) Zoned pyroxenes and/or zoned garnets from Koidu eclogites. The various fields are from Figure 9a. Sample K91-5 is apatite-bearing. (c) Mantle and crustal eclogites and related rocks. The sources and localities are the same as in Figure 2d.

Amphibole

Amphiboles in the Koidu eclogites are secondary and occur either as euhedral to subhedral grains (1–2 mm) or as interstitial grains (<1 mm) along grain boundaries and cracks. Extremely fine grained amphiboles may also be present as part of kelyphite-like assemblages replacing garnet. Two types of amphibole have been identified in the Koidu eclogites: a greenish-brown calcic-amphibole and a rare lavender-blue alkali variety; both types of amphibole are commonly associated with phlogopite as metasomatic phases.

In thin section, calcic-amphiboles are strongly pleochroic, ranging from light green to dark green and pale brown to dark brown in color. Average compositions based on amphiboles in eight eclogites (Table A9) show a small variation in weight percent CaO (9.81–12.22), MgO (10.1–13.9), and TiO_2 (0.99–2.64). The compositions of amphibole along garnet and pyroxene grain boundaries and those in the moat between the core and rim of zoned garnets are indistinguishable. According to the classification scheme established by *Leake* [1978], amphibole in one of these eclogites is categorized as pargasite, whereas those in the remaining seven are ferroan pargasites. Pargasite has been recognized as the preeminent phase in patently metasomatized peridotite xenoliths from alkali basalts [*Dawson and Smith*, 1982]. Other amphibole varieties reported from the Koidu eclogites include pargasitic and edenitic hornblende [*Hills and Haggerty*, 1989]. Edenitic hornblende has also been reported in eclogites from Orapa [*Tollo*, 1982], and primary kaersutite has been recognized in the Stockdale eclogites [*Meyer and Brookins*, 1971].

Subhedral to anhedral, pleochroic lavender-blue to light brown alkali amphibole has been identified in seven eclogites. The amphibole is fibrous in some samples and occurs mainly in contact with calcic-amphibole. Chemical variations of the alkali amphiboles (Table A10) among samples are significant with respect to weight percent TiO_2 (0.52–3.55), MgO (13.4–17.4), and FeO (8.0–16.9) but are fairly constant in weight

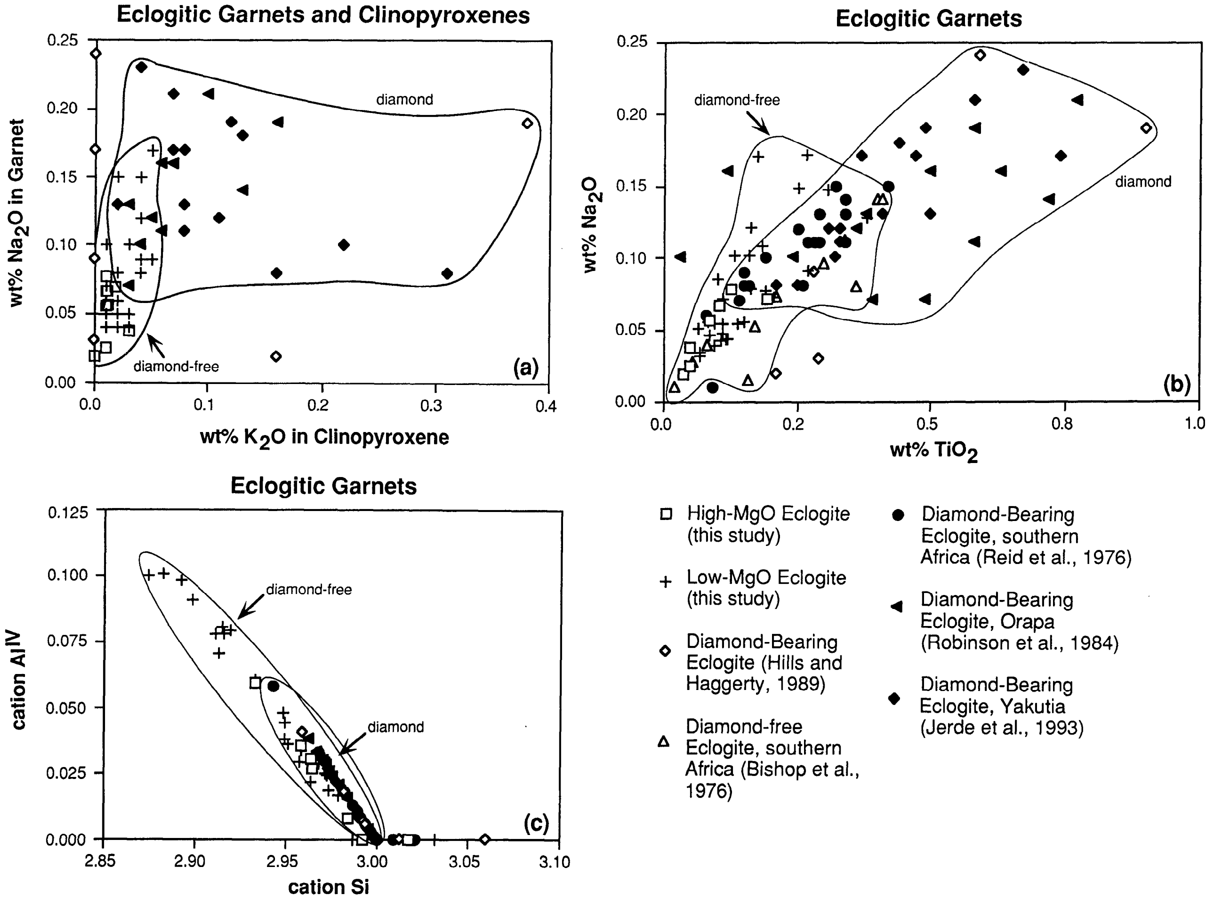


Figure 10. Plots of (a) weight percent K_2O in clinopyroxene versus Na_2O in garnet, (b) weight percent TiO_2 versus Na_2O in garnet, and (c) cation Si versus cation ^{IV}Al in garnet for diamond-bearing and diamond-free eclogites.

percent CaO (0.88–1.93), Na_2O (8.11–9.12), and K_2O (0.81–1.43). Amphibole analyzed from two of the four eclogites may be classified as eckermannite; those in the other two are grouped as magnesio-arfvedsonite [*Leake*, 1978].

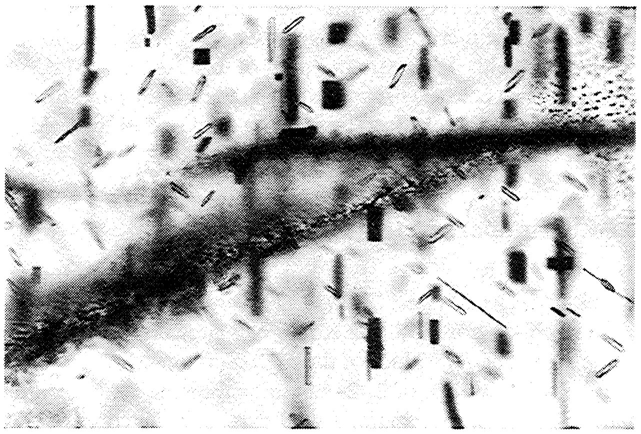


Figure 11. Photomicrograph of oriented apatite rods (colorless) and rutile plates (black) in the core of a pyroxene from Koidu eclogite KEC-81-12. Transmitted plane polarized light. Width of photomicrograph is 0.69 mm.

Phlogopite

Secondary phlogopite is abundant in some eclogites and may account for up to 10 modal percent. The majority of phlogopite occurs in clusters with amphibole and typically surrounds kelyphitized garnet. Phlogopite is strongly pleochroic from colorless to dark brown in thin section. It shows hexagonal or rectangular forms with lengths of 0.5–2.0 mm. Average compositional data for this metasomatic phase are given in Table A11. In general, phlogopites in high-MgO eclogites contain higher weight percent MgO (24.7–24.9), lower FeO (3.82–4.61), and lower TiO_2 (0.35–0.65) than the low-MgO varieties (MgO = 18.9–23.3 wt %, FeO = 6.52–8.47 wt %, TiO_2 = 0.34–3.21 wt %). The Cr_2O_3 contents displayed by the phlogopites (<0.46 wt %) are low compared with most peridotitic phlogopites, which may contain up to 2.12 wt % Cr_2O_3 [*Dawson*, 1980; *Field et al.*, 1989]. Similar abundances of weight percent K_2O (9.20–10.30) and BaO (<0.05–0.82) have been reported in other eclogitic and peridotitic phlogopites [*Field et al.*, 1989; *Hills and Haggerty*, 1989].

Replacement of phlogopite by chlorite is identified in three Koidu eclogites (K91-7, K91-26, and K91-61). Phlogopite-chlorite intergrowths have been described in the Koidu kimberlite [*Tompkins et al.*, 1984] and in eclogites from Roberts Victor [*Kushiro and Aoki*, 1968]. The petrographic and compositional characteristics of phlogopites with intergrowths do not differ from those free of intergrowths. Chlorite shows strong pleochroism from greenish-blue to light brown and preferentially replaces phlogopite along cleavage planes. Chlorite in this study (Table A12) has higher weight percent Al_2O_3 (14.6–17.5) and lower weight percent SiO_2 (30.5–31.2) and FeO (7.4–14.3) than the discrete chlorite xenoliths from the Koidu kimberlites [*Tompkins et al.*, 1984].

Rutile and Ilmenite

Rutile with fine-grained ilmenite intergrowths is the most common accessory phase in the eclogites. Depending on the abundance of ilmenite lamellae, rutile grains vary from clear, brownish-yellow to dark brown or opaque in grains saturated in ilmenite. Rutile (Table A13) in this study (94.9–99.9 wt % TiO_2) is diluted by small amounts of FeO (0.35–3.36 wt %). The compositions of rutiles are almost identical to other eclogitic rutiles [*Meyer and Boctor*, 1975; *Smith and Dawson*, 1975] but are low in Cr_2O_3 (0.05–0.32 wt %) compared to peridotitic rutiles which may contain up to 7.2 wt % Cr_2O_3 [*Smith and Dawson*, 1975]. Rutiles in peridotitic metasomites [*Haggerty*, 1991] are distinctive with high weight percent Cr_2O_3 (8.21), FeO (4.1), and Nb_2O_5 (20.9).

Ilmenite in host rutile is too fine grained for microprobe analysis. Discrete ilmenite grains are rare but are present in five Koidu eclogites. Compositions of ilmenite are listed in Table A14. TiO_2 is relatively constant (47.8–52.6 wt %), but the abundances of weight percent FeO (28.0–36.4), Fe_2O_3 (7.07–13.58), and MgO (4.60–10.57) differ from rock to rock. As is the case of rutile, discrete ilmenite from Koidu is compositionally similar to some ilmenites from other mantle eclogites and peridotites [*Meyer and Boctor*, 1975; *Smith and Dawson*, 1975; *Haggerty*, 1991].

Acmite and Feldspar

Long thin needles of acmite are recognized in albite and sanidine assemblages (150–250 μm in diameter) which are present between grain boundaries in four Koidu eclogites (K91-18, K91-22, K91-23, and K91-36). Acmite is colorless to slightly blue and generally ranges from 10 to 100 μm in length. Feldspar crystals are colorless and anhedral. Albite is the dominant phase with small patches of associated sanidine. Average compositions for acmite and feldspar are given in Tables A15 and A16. The association of the acmite-feldspar assemblage with secondary pargasitic amphibole suggests a complex metasomatic origin for these unusual assemblages, involving the introduction of Na, K, and Fe or the extraction of these elements from primary garnet and clinopyroxene.

Reconstructed Bulk-Rock Compositions

Modal abundances and mineral compositions of garnet and pyroxene were used to compute estimated bulk-rock compositions for the eclogites. The reliability of this method has been tested by comparing the bulk Koidu eclogite compositions [*Hills and Haggerty*, 1989] previously obtained by X ray fluorescence with compositions reconstructed from the same eclogite suite. The analyzed and reconstructed bulk MgO, FeO, and CaO of these rocks are almost identical, as indicated by the 1:1 linearity in plots of analyzed versus reconstructed weight percent MgO, FeO, and CaO (Figures 12a–12c). Slight deviation between these compositions may be explained by inexact modal estimates or small analytical uncertainties. Prominent outliers in the diagrams, however, are mostly due to late stage alteration which may be up to 20 modal % in some samples [*Hills and Haggerty*, 1989]. The horizontal distance between an outlier and the 1:1 line represents the amount of a specific component being added to (positive displacement from the 1:1 line) or removed from an eclogite (negative displacement from the 1:1 line). Samples that are altered or metasomatized but fall near the 1:1 line imply that only some elements have been affected. For example, sample KEC-81-18 is an outlier in Figure 12a but not in Figures 12b and 12c, suggesting that the bulk MgO content of this rock might have been enriched during metasomatism but the effects on FeO and CaO are insignificant. Slight variations between analyzed and calculated SiO_2 and Al_2O_3 can be explained by metasomatism and the presence of other secondary silicate minerals, such as amphibole and phlogopite. It is emphasized that only garnet and pyroxene are involved in the computation in order to obtain the premetasomatic composition. The amount of calculated P_2O_5, TiO_2, and K_2O may be underestimated, as the major hosts of these components (apatite, rutile, and phlogopite) have not been included. Nonetheless, the modal recombination method appears to be reliable for assessing bulk major element composition and serves at least as a qualitative estimate of minor element composition.

The reconstructed bulk compositions of 32 Koidu eclogites in this study are listed in Table A17. Bulk compositions of eclogites with zoned garnets and/or zoned pyroxenes have not been determined. On the basis of bulk MgO contents the eclogite suite may be divided into two groups. High-MgO eclogites (15.04–18.85 wt %) are characterized by pyrope and diopside, whereas low-MgO eclogites (7.10–12.21 wt %) are dominated by grossular and almandine and jadeitic-diopside and omphacite. In general, high-MgO eclogites have lower weight percent CaO (8.64–13.97) and FeO (5.60–8.60) than low-MgO rocks, which contain 9.94–16.31 wt % CaO and 8.99–17.00 wt % FeO. The reconstructed concentration of major elements in these eclogites is similar to mantle eclogites from kimberlite localities worldwide [*Dawson*, 1980]. The major elements of these rocks broadly resemble basalts, picrites, and komatiites, except for the low TiO_2 and K_2O contents which result from the exclusion of rutile and secondary phlogopite in bulk-rock computations. Bulk compositions of the eclogites plotted in a CaO-MgO-FeO ternary (Figure 13a) show that almost all eclogites fall into the bimineralic eclogite and peraluminous eclogite fields compiled by *Dawson* [1980]. Approximately two thirds of the eclogite population lie in the field of basalts, picrites, and komatiites, with none lying in the grospydite field. In a ternary plot of CaO-Al_2O_3-(MgO + FeO), the two eclogite groups overlap (Figure 13b) and form a tight cluster in the bimineralic eclogite field and in the field of basalts, picrites, and komatiites [*Dawson*, 1980].

Geothermobarometry

K_D Values

The $K_D^{\text{Fe-Mg}}$ [*Ellis and Green*, 1979] values of coexisting garnet and clinopyroxene of 32 Koidu eclogites are presented in Table A18. The K_D values of the high- and low-MgO eclogites overlap, with a range of 2.42–6.38. For the nine xenoliths that contain zoned garnet and/or zoned pyroxene the cores of garnet and pyroxene are used to calculate K_D for core compositions. Corresponding K_D for rim compositions are based on the average compositions of garnet rims and the pyroxene rims. With the exception of specimens K91-18 and K91-64 the

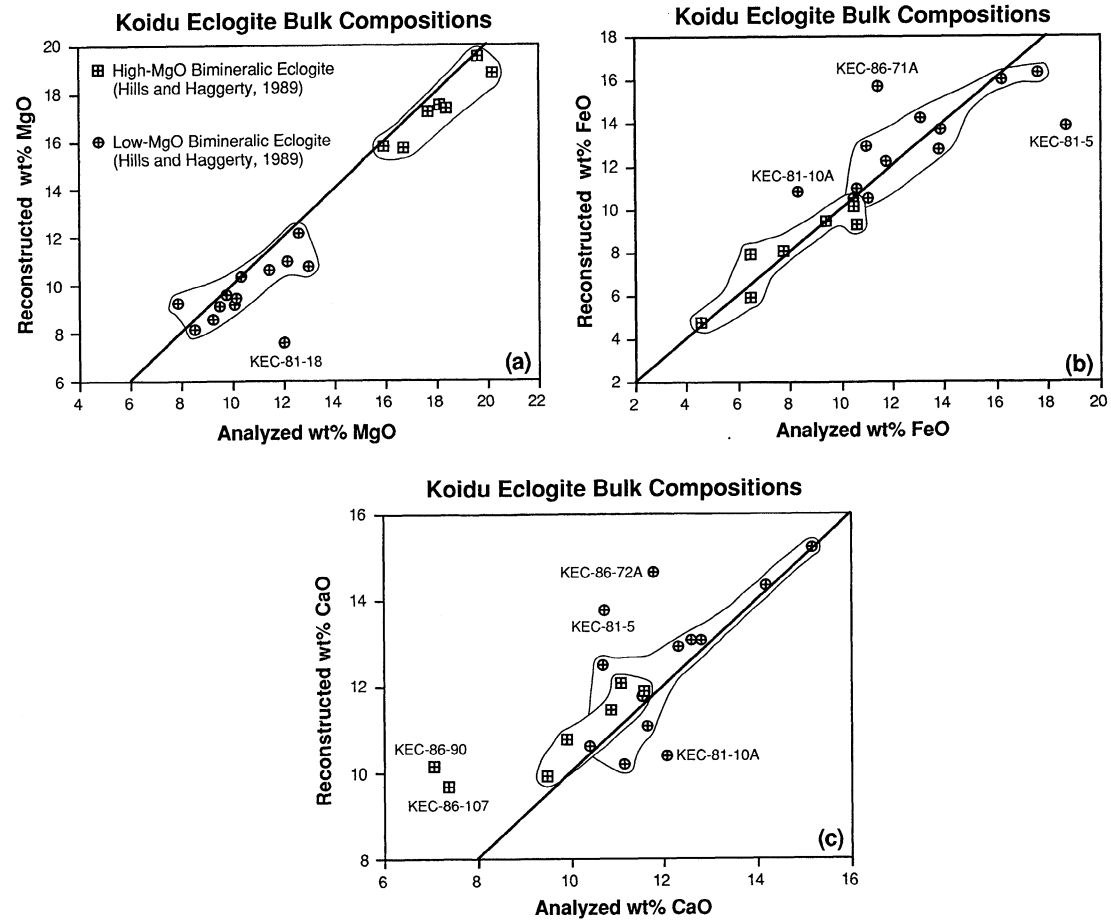


Figure 12. Analyzed versus reconstructed bulk compositions for Koidu eclogites. (a) Weight percent MgO, (b) weight percent FeO, and (c) weight percent CaO. The samples labeled are extensively altered. Data are from *Hills and Haggerty* [1989].

K_D of core compositions (2.73–6.17) are higher than coexisting rims (2.49–5.50) in individual xenoliths (Table A19). The range of K_D in these eclogites is comparable to those reported from other kimberlitic eclogites [*Kushiro and Aoki*, 1968; *Bishop et al.*, 1978; *Dawson and Carswell*, 1990, and references therein].

A positive curvilinear correlation between mole percent grossular and ln K_D for the Koidu eclogites is present in Figure

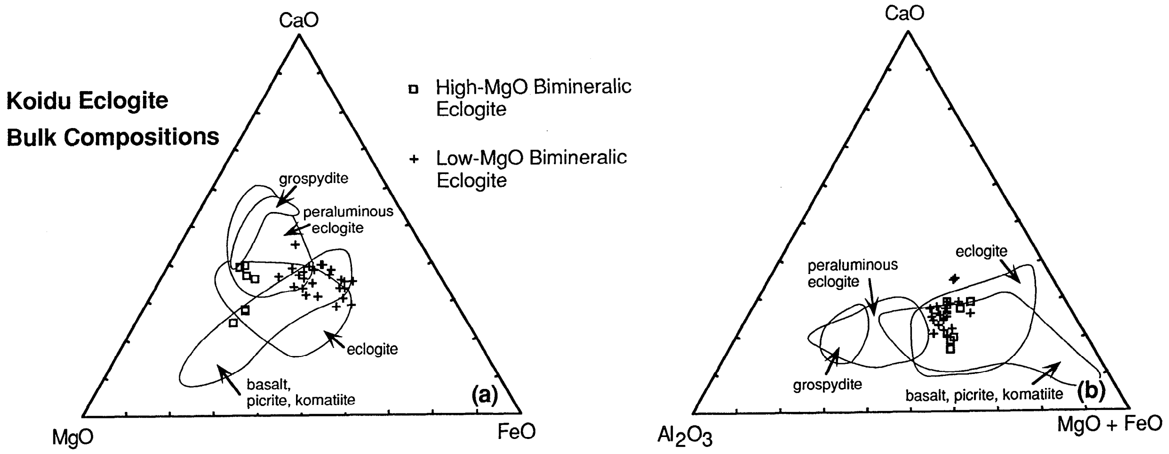
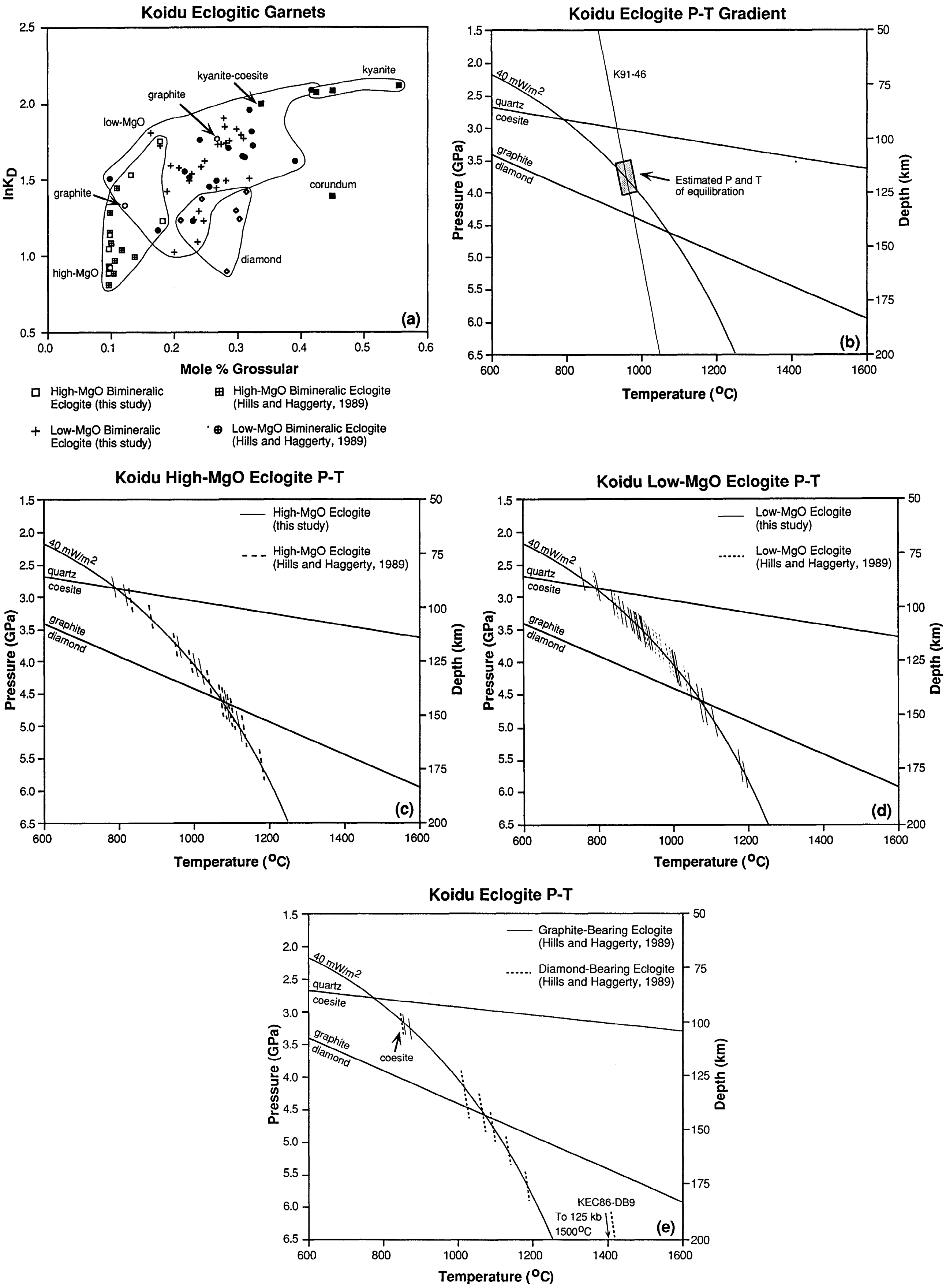


Figure 13. Ternary diagram of reconstructed bulk rock (a) weight percent CaO-MgO-FeO and (b) weight percent CaO-Al_2O_3-(MgO + FeO) for Koidu eclogites. The fields are outlined and labeled after *Dawson* [1980].


Koidu Eclogitic Garnets
lnK_D
Mole % Grossular
kyanite-coesite
kyanite
graphite
low-MgO
graphite
corundum
diamond
high-MgO
(a)
□ High-MgO Bimineralic Eclogite (this study)
⊞ High-MgO Bimineralic Eclogite (Hills and Haggerty, 1989)
+ Low-MgO Bimineralic Eclogite (this study)
• ⊕ Low-MgO Bimineralic Eclogite (Hills and Haggerty, 1989)
Koidu Eclogite P-T Gradient
Pressure (GPa)
Depth (km)
Temperature (°C)
40 mW/m2
quartz
coesite
graphite
diamond
K91-46
Estimated P and T of equilibration
(b)
Koidu High-MgO Eclogite P-T
High-MgO Eclogite (this study)
High-MgO Eclogite (Hills and Haggerty, 1989)
(c)
Koidu Low-MgO Eclogite P-T
Low-MgO Eclogite (this study)
Low-MgO Eclogite (Hills and Haggerty, 1989)
(d)
Koidu Eclogite P-T
Graphite-Bearing Eclogite (Hills and Haggerty, 1989)
Diamond-Bearing Eclogite (Hills and Haggerty, 1989)
coesite
KEC86-DB9
To 125 kb
1500°C
(e)

14a. Also plotted for comparison are eclogite data from *Hills and Haggerty* [1989]. High-MgO eclogites in both studies yield lower K_D and grossular content than low-MgO varieties. The corundum-bearing eclogite in the *Hills and Haggerty* [1989] study is an outlier, with ln K_D = 1.40 and mole percent grossular is 0.45. Both the diamond-bearing and kyanite-bearing eclogites [*Hills and Haggerty*, 1989] form a separate subgroup from the bimineralic eclogites.

Temperatures and Pressures of Equilibration

The equilibration temperatures for the eclogites have been calculated at an assumed upper mantle pressure range of 1.5–6.5 GPa, using the *Ellis and Green* [1979] geothermometer. At 4.5 GPa, temperatures obtained by the *Powell* [1985] method are 10°–25°C lower than those of *Ellis and Green* [1979], whereas the differences between *Krogh* [1988] and *Ellis and Green* [1979] may be up to ±100°C, especially for eclogites with low-Ca garnets (Tables A18 and A19). Preference is given here to temperature calculated from *Ellis and Green* [1979] to be consistent with the method used for other Koidu eclogites [*Hills and Haggerty*, 1989].

In the absence of appropriate geobarometers for bimineralic garnet + clinopyroxene rocks the equilibration pressures of the eclogites can be estimated if it is assumed that P is related to T and that both are constrained by reference to a typical Precambrian shield geotherm of 40 mW/m^2 [*Pollack and Chapman*, 1977]. This geotherm is in accord with surface heat flow from the Man Shield in the West Africa craton [*Brigaud et al.*, 1985]. The PT gradients at 1.5–6.5 GPa for the xenoliths are plotted on this reference geotherm [*Pollack and Chapman*, 1977], and the equilibration P and T are obtained at the points of intersection between the geotherm and individual PT gradients (Tables A18 and A19), as illustrated in Figure 14b for sample K91-46. The eclogites in this study show a large spread in the estimated PT, extending from 760°C at 2.8 GPa to 1188°C at 5.8 GPa (Table A17 and Figures 14c and 14d), which coincides with depths of about 90–180 km [*Dziewonski and Anderson*, 1981]. Most of the eclogites fall into the graphite stability field, but a few high-T (>1100°C) eclogites with no observable diamonds plot in the diamond stability field [*Kennedy and Kennedy*, 1976]. Except for samples K91-18 and K91-64, eclogites with zoned garnet and/or pyroxene show lower PT for core compositions than that of rim compositions (Table A19).

The equilibration PT of a coesite-bearing eclogite, two graphite-bearing eclogites, and six diamond-bearing eclogites in *Hills and Haggerty*'s [1989] study is plotted in Figure 14e. The graphite-bearing samples lie well within the graphite stability field, and the diamond-bearing samples lie in or close to the diamond stability field [*Kennedy and Kennedy*, 1976]. Sample KEC86-DB9 contains both diamond and graphite and displays equilibration PT in the transition zone. The coesite-bearing eclogite also falls in the coesite stability field [*Bohlen and Boettcher*, 1982]. The estimated PT of these accessory mineral-bearing eclogites is consistent with the observed presence of diamond, graphite, or coesite and justifies the method of PT derivation as a useful approximation of equilibration conditions.

Figure 15 shows histograms of equilibration PT for the eclogites included in this study and those by *Hills and Haggerty* [1989]. Although there is an overlap in both the P and T range for high- and low-MgO eclogites, a high proportion of high-MgO eclogites peaks at 1080°C at 4.7 GPa (~150 km deep) to 1130°C at 5.2 GPa (~165 km deep), whereas the low-MgO eclogites cluster at 880°C at 3.3 GPa (~105 km deep) to 930°C at 3.8 GPa (~120 km deep). Statistical analyses show that the PT of these two eclogite groups differ significantly. Assuming that eclogites equilibrated at the site of derivation, the PT gradients in Figure 15a and 15b imply that these xenoliths may have sampled at least two regions in the upper mantle. On the basis of the higher equilibration PT and the absence of diamond, *Hills and Haggerty* [1989] proposed that high-MgO eclogites formed in the deeper, fertile, more oxidized, and highly convective asthenosphere in which carbon is present as a volatile phase (e.g., CO and CO_2). Low-MgO eclogites (bimineralic or with accessory minerals) in general have lower equilibration PT and indicate an origin in the shallower, depleted, and less oxidized lithosphere where free elemental carbon exists. The estimated PT of eclogites in this study are in accord with the above conclusions made by *Hills and Haggerty* [1989]. It is noted, however, that the distribution of high- and low-MgO eclogites may be more complicated because *Reid et al.* [1976] show that both high-MgO (~1750°C) and low-MgO (~1050°C) eclogites from southern Africa may be diamondiferous, and therefore some high-MgO eclogites may or may not originate in the oxidized asthenosphere. Whether carbonaceous eclogites contain diamond or graphite will depend on the position of the eclogite body relative to the diamond-graphite stability region, on the geotherm of the upper mantle, on redox, and on the rate of xenolith transport. Diamondiferous eclogites may require very special conditions for formation and preservation [*Haggerty*, 1986].

Origin of Bimineralic Eclogites

The intrusions at Koidu are type I kimberlites [*Smith*, 1983], dominated by olivine and with abundant peridotitic and eclogitic suite diamonds. In conformity with other subcontinental, subcratonic lithospheres and asthenospheres we assume that the West Africa craton is dominated by garnet lherzolite, garnet harzburgite, and dunite; other rock types, including eclogite, are garnet pyroxenite, wehrlite, websterite, and discrete megacrystic xenoliths (olivine, clinopyroxene, orthopyroxene, garnet, ilmenite, and phlogopite); all of which (with the possible exception of eclogite) are generally regarded as being magmatic [*Nixon*, 1987]. From bulk compositions the eclogite suite at Koidu spans the range from basalt through picrite to komatiite which in accord with modern and ancient analogs would require a superposition of or an interdigitated sequence of Archean oceanic (basalt) and continental (komatiite) crusts, if subduction and recycling are invoked for the

Figure 14. (Opposite) (a) Plot of mole percent grossular versus ln K_D for Koidu eclogites. Accessory mineral-bearing eclogites are from *Hills and Haggerty* [1989]. (b) PT gradient for eclogite K91-46. The equilibration PT is obtained at the intersection between the PT gradient and the 40 mW/m^2 Precambrian shield geotherm [*Pollack and Chapman*, 1977]. The diamond-graphite stability curve is from *Kennedy and Kennedy* [1976]. The coesite-quartz stability curve is from *Bohlen and Boettcher* [1982]. Corresponding depths for pressures are from *Dziewonski and Anderson* [1981]. Plots of equilibration temperatures versus pressures for (c) high- and (d) low-MgO eclogites and (e) for low-MgO coesite-bearing, graphite-bearing, and diamond-bearing eclogites. Sample KEC86-DB9 contains both diamond and graphite.

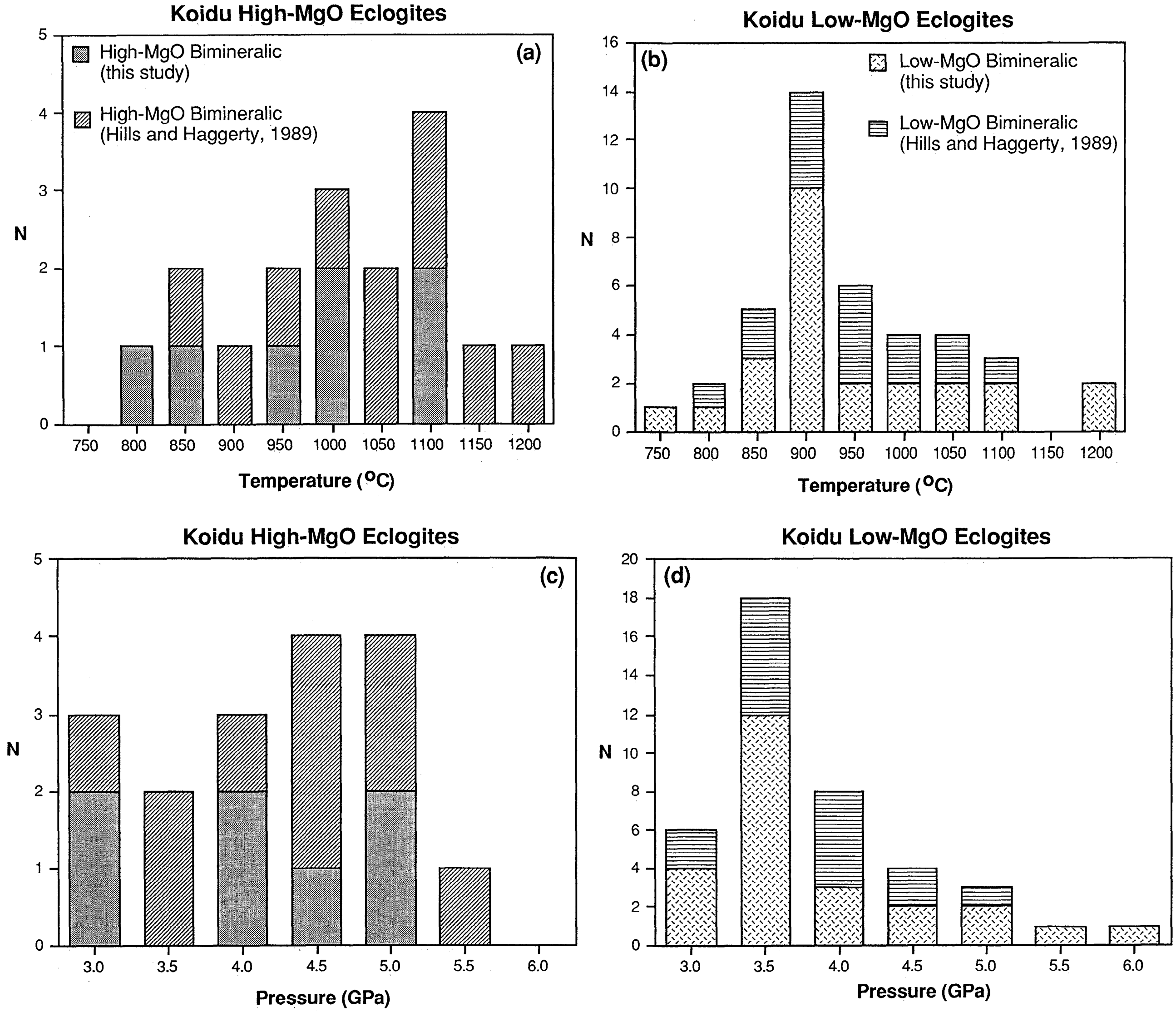


Figure 15. Histograms of equilibration temperatures for Koidu (a) high- and (b) low-MgO eclogites and histograms of equilibration pressures for Koidu (c) high- and (d) low-MgO eclogites.

origin of eclogites. This is not an impossible but is a most unlikely scenario given the contrasts in buoyancy and the complexity of tectonism that would be required at a single site. The origin of the bimineralic suite of eclogites at Koidu is equivocal, but we concur with the view of *MacGregor and Carter* [1970], *Smyth and Caporuscio* [1984], and *Taylor and Neal* [1989] that the cumulate-like textures displayed by some Koidu eclogites may be indicative of an igneous origin. *Rudnick et al.* [1993] linked eclogites having high-Na_2O clinopyroxenes to Na_2O-rich protoliths and argued that metamorphosed subducted oceanic crust (<3 wt % Na_2O) is an unlikely protolith because it requires >3.2 wt % Na_2O to crystallize omphacitic pyroxene; *Rudnick and Green* [1994], however, propose a more complex model of subduction and partial melting. The majority of Koidu eclogitic pyroxene contains 3.5–7.2 wt % Na_2O. In our model we assume that the protolith is garnet lherzolite, enriched by alkali silicate melt metasomatism. High PT experiments on lherzolite melting show that garnet + clinopyroxene are early melt products [e.g., *Mysen and Kushiro*, 1977]. If these melts are mobilized from the source but crystallized at depth, the result would be eclogite. A heat source for melting is required at depths exceeding the graphite/diamond stability limit (i.e., >180 km at ~1300°C). One possibility is incubation beneath a supercontinent [*Anderson*, 1994] and the other is by plumes from the D″ core-mantle boundary layer [*Haggerty*, 1994]. In both thermal models the range of eclogite compositions at Koidu can possibly be accounted for by crystallization from progressively evolved liquids. In the plume model, however, the distribution in and subsequent sampling of eclogites of diverse compositions along a column can be explained by differences in temperature and degrees of partial melting in accord with the model by *Campbell et al.* [1989, p. 697], "... heads it's basalts, tails it's komatiites." This is furthermore consistent with low-MgO eclogites having lower temperatures of equilibration and high-MgO eclogites having higher temperatures of equilibration. Compositional trends within each eclogite group are also consistent with an expected range of melting within the plume head and the tail.

From the abundance of eclogite xenoliths at Koidu we assume that eclogite is a prominent rock type at subcratonic depths, but reasons for the absence (none located) or extreme rarity of peridotites at this and other localities are uncertain. One possibility is preferential sampling by the kimberlite(s), and the other is silico-carbonatitic or carbonatitic metasomatism [*Green and Wallace*, 1988] (in which olivine + SiO_2 + $CaCO_3$ = clinopyroxene + CO_2, and orthopyroxene + $CaCO_3$ = clinopyroxene + CO_2) which appears to be widespread in continental and oceanic upper mantle settings [e.g., *Hauri et al.*, 1993]; natro-carbonatitic metasomatism should precipitate diopside-jadeite (omphacite). This style of metasomatism, which is pervasive in the Koidu kimberlite [*Tompkins and Haggerty*, 1984], would liberate peridotitic suite diamonds and may even be responsible for the genesis of some eclogites, with the addition of melting and through the removal of olivine and orthopyroxene from four-phase garnet lherzolite.

A third possibility for the origin of bimineralic eclogites is exsolution of clinopyroxene from majoritic garnet. A candidate is eclogite KEC86-DB9 (Figure 14e) which equilibrated at 12.5 GPa and 1500°C (350–400 km), and another eclogite (KEC86-8) with minor orthopyroxene [*Hills and Haggerty*, 1989] is remarkably similar in mineral chemistry to majoritic garnet associations from the Kaapvaal Craton [*Haggerty and Sautter*, 1990]. Some eclogites may therefore have originated at or within the transition zone.

In conclusion, eclogites appear to be distributed throughout the upper mantle and transition zone and eclogites possibly have multiple origins.

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A. T. Fung, C. F. Mineral Research, Ltd., 1677 Powick Road, Kelowna, British Columbia, Canada V1X 4L1. (e-mail: CF_Minerals@cyberstore.ca)

S. E. Haggerty, (corresponding author) Morrill Science Center, Geology Department, University of Massachusetts, Amherst, MA 01002. (e-mail: haggerty@eclogite.geo.umass.edu)

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